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ATMOSPHERIC ELECTRICITY

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O. W. RICHARDSON

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ATMOSPHERIC ELECTRICITY

CHAPTER I

THE IONISATION OF THE ATMOSPHERE

THE electrical conductivity of the air, first established by Linss in the year 1887, plays an important part in all the phenomena included under the title Atmospheric Electricity. Since 1901, when Elster and Geitel and C. T. R. Wilson independently discovered that this conductivity is due to the presence of small carriers of positive and negative charge called ions, the nature and the origin of these ions have been the subject of many investigations. Two different paths have been followed, the one dealing with the number and the nature of the atmospheric ions, the other with the radiations producing them. *The two paths converge on the problem of the* ionisation-balance of the atmosphere, the manner in which an equilibrium is maintained between the rate of creation and the rate of disappearance of the ions. It is only of recent years that an answer to this problem has emerged.

The investigation of the ionisation of the air has been rendered exceptionally difficult by the number of separate factors involved and by the variation of each with local geological and meteorological conditions. Much remains to be done before the conditions governing the conductivity of even the lowest layers of the atmosphere are completely understood; observations at greater heights in the regions accessible to balloons and airships are just emerging from the difficult pioneer stage, and the very important regions of the upper air which lie

beyond fifteen kilometres above the ground can at present be examined only indirectly by deductions based upon the mode of propagation of upwardly directed wireless waves and upon the variations of the earth's magnetic field.

§ 1. The Atmospheric Ions, Small and Large

A gas molecule is ionised when some external agency supplies the energy necessary to remove an electron from it, whereupon the positively charged residue of the molecule and the ejected electron, both of which under ordinary pressures very soon attach themselves to one or more uncharged molecules, form what is called an ion-pair. Such ions are known as "normal," "small," or "fast," and when found in the atmosphere have mobilities of the same order as those determined for pure dry air in laboratory experiments, about 1·5 cm./sec. in a field of gradient 1 volt per cm.

The presence in the air of minute particles of dust, of the products of combustion, and of extremely small drops of water even when the air is unsaturated, leads very often to the capture of a normal ion by one of these condensation or Aitken nuclei, as they are termed, and so to the formation of another type of ion of low mobility, ranging from ·0005 to ·0003 cm./sec., known as "large," "slow," or "Langevin" ions. Roughly speaking, one-third of the condensation nuclei present in the air is electrically neutral, the other two-thirds carry, nearly equally, positive and negative charges captured from small ions. The interplay between the large and the small ions and the uncharged nuclei is of considerable importance.

Besides these two main types of atmospheric ion, a group of "intermediate" ions whose mobilities lie between ·1 and ·01 cm./sec. has been found to be present under special meteorological conditions such as low relative humidity. While the origin of this type of ion is at present uncertain, its existence emphasises the fact

that the customary division into the two main classes, large and small, requires refinement, and future work will no doubt proceed in the direction of dividing the ions according to their mobilities into a larger number of groups. J. Nolan and de Sachy have recently found evidence of such groups in the small and the large ion "families." [1]

It has been found that the great majority of the ions, large and small, carry a single elementary charge, either positive or negative.[2] Over land areas the large ions considerably outnumber the small ones, often in the ratio of ten to one. Thus while the numbers, n_+ and n_-, of small ions of each sign per c.c. of the air over land usually range from 300 to 1000, those of the large ions, N_+ and N_-, amount to from 1,000 to 80,000 per c.c. Since practically all the large ions are formed from the capture of small ones by nuclei, an increase in the number of these nuclei, such as is due to the fires of an industrial area, will increase the proportion of the large ions in the air at the expense of the small ions.

Over the oceans this state of affairs is reversed ; the air away from land areas is but poorly supplied with nuclei and the large ions are in the minority. The mean values of n_+ and n_- at sea lie between 500 and 700 per c.c., while N_+ and N_- are about 200 per c.c. In view of their comparative freedom from capture by nuclei, it is at first sight rather surprising that the number of small ions per c.c. in mid-ocean differs but little from the same quantity over the land, where the average values of n_+ and n_- are also about 600 per c.c. This is, however, to be explained (§ 12) by the fact that the decreased rate of disappearance of the small ions over the sea is balanced by a decrease in the rate at which they are produced.

§ 2. The Determination of the Number of Ions per c.c.

To determine the numbers of the small ions, positive and negative, per c.c. (n_+ and n_-) an arrangement known

as an Ebert ion-counter is usually employed. This consists in essentials of a long earthed metal tube A (Fig. 1) on the axis of which is mounted a charged insulated rod B, connected to a quartz fibre electroscope. Air is sucked through the tube for about five minutes by means of a clock-work driven fan F and the speed of the air-stream is arranged to be slow enough for all the small ions entering A with charges opposite in sign to that on the central rod to be drawn to B before reaching the bottom of the tube. The speed of flow must at the same time be great enough to make it impossible for any appreciable number of large ions to have time to reach the rod. If v is the volume of the air sucked through in the course of the measurement (indicated by an anemometer C registering its revolutions on a dial), the quantity of electricity gathered in by the central electroscope system is ven_-, where e is the elementary charge on a single ion and the rod is supposed to have been positively charged. If then C is the capacity of the whole central system, rod and electroscope, and V and V′ are the potentials indicated by the electroscope before and after the experiment, $C(V - V') = ven_-$, an equation from which n_- can be

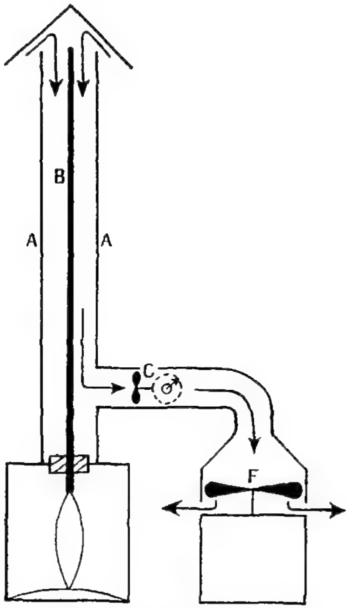

FIG. 1.—Ebert ion-counter.

found. To determine n_+, the measurement must be repeated with the rod B charged negatively.

The same method, using a longer tube, a stronger field in its interior, and a slower draught of air, and thus giving time for the capture of all the slow ions as well, can be applied to the determination of $N + n$. A combination of the two determinations then yields N_+ and N_-, the numbers of the large ions per c.c. of each sign.

The values found for n_+ and n_- are not usually the same; the ratio n_+/n_- over both land and sea has a mean value of 1·22. Although, as explained in § 30, the earth's electric field may sometimes affect the apparatus in such a way as to give a spuriously high ratio, Swann, who has made a special study of the point, concludes that when all precautions are taken a real excess of positive ions must be admitted. Probably the main factor causing this excess is the superior mobility of the small negative ion; Nolan has found that k_-/k_+ has the value 1·16. A lower mobility implies a lower velocity of thermal agitation for the positive than the negative ion, and consequently less likelihood of capture by condensation nuclei. Another contributing cause may be the Ebert effect discussed in § 38. This general excess of positive small ions does not necessarily involve a nett positive charge in the air; it is easily compensated for by a comparatively small excess in the much larger number of large negative ions.

An instrument known as the Aitken dust-counter is used to find the total number of nuclei, charged and uncharged, per c.c. It consists of a small expansion chamber filled with a sample of the air and saturated with water-vapour. The sudden movement of a piston gives an adiabatic expansion which causes drops of water to condense upon the nuclei; the resulting cloud falls upon a glass plate divided into squares and the number of droplets is counted under a microscope. Though called a dust-counter it does not record the coarser dust in the air.[3]

§ 3. The Polar Conductivities and the Ionic Mobilities

A simple relation holds between the conductivity of the air and the numbers per c.c. and the mobilities of the ions responsible for conduction. Consider a point in the atmosphere at which the electric field intensity is F and the specific conductivity, the reciprocal of the specific resistance, is λ. The conduction current, i, through unit area drawn perpendicular to the field is then equal to $F\lambda$, for F is numerically equal to the potential difference, and $1/\lambda$ is the resistance, between the ends of a unit cube of the air. In this equation we assume that conditions are such that Ohm's law can be applied to the air. Since the current is carried by positive and negative streams of ions moving in opposite directions we may write

$$i = F(\lambda_+ + \lambda_-), \quad . \quad . \quad . \quad (1)$$

where λ_+ and λ_- are called the polar conductivities.

Actually the existence of the two main types of ion, large and small, leads to the flow of four different ion-streams, two up and two down if the field is vertical, in each of which the velocity is different, being given by $F \times k$, where k is the corresponding mobility. The total current is thus

$$i = Fe(n_+k_+ + n_-k_- + N_+K_+ + N_-K_-), \quad . \quad (2)$$

where k_+ and k_-, K_+ and K_-, are the mobilities of the positive and the negative ions, large and small respectively, and e is the elementary charge. From (1) and (2), separating out the oppositely moving streams,

$$\lambda_+ = (k_+n_+ + K_+N_+) \text{ and } \lambda_- = (k_-n_- + K_-N_-). \quad (3)$$

In clear country air over land, n_+ and n_- number about 600 per c.c., while N_+ and N_- are about 2000 per c.c.; k_+ and k_- are 1·5 approximately and K_+ and K_-, ·0004. Here, therefore, the ratio of the first to the second term in each bracket above is 1100 : 1 and most

of the conductivity of the air is due to the small ions. Near large towns, however, n is usually of the order of 100 per c.c. and N, on account of the heavy pollution of the atmosphere may be 50,000. Under these conditions the second term would be one-seventh of the first and the large ions would play a part in the transfer of electricity through the air.*

It is not usual to find the polar conductivities from equations (3) above, but directly, from measurements of the rate which the charge on an insulated body exposed to the free air is dissipated by the flow of ions to it. Consider a negatively charged body exposed to moving air and let σ be the density of the electrification on any portion of area dS.[1] Positive ions will move to the body, and since the field F just outside the surface is $4\pi\sigma$, the current per unit area tending to dissipate the charge will be $F\lambda_{+} = 4\pi\sigma\lambda_{+}$. The total dissipation current, obtained by integrating over the whole exposed surface, is $4\pi\lambda_{+}\int\int\sigma dS$ or $4\pi\lambda_{+}Q$, where Q is the charge exposed to the air. Since this current represents the rate at which the body loses charge, we have $-dQ/dt = 4\pi\lambda_{+}Q$, a relation which provides a means of determining λ_{+}. A similar equation, with λ_{-} substituted for λ_{+}, holds for the case of a positively charged body.

An arrangement used for this purpose is due to Gerdien (Fig. 2); it resembles the Ebert apparatus already described, but the air-stream is now made so rapid and the field in the interior of the tube so weak, that only a very small fraction of the total number of ions passing through A has time to reach the central rod. In this case the numbers of ions per c.c. are not appreciably altered by the experiment and Ohm's Law applies.

If the potential V of the negatively charged central electroscope system is observed to rise at a rate dV/dt and if C is the total capacity of the central system, rod and electroscope, we have $-dQ/dt = -CdV/dt$. Also,

* The methods used to determine ionic mobilities are described in text-books on the conduction of electricity through gases

if C′ is the capacity of that portion of the central system exposed to the stream of air, the rod B and its support, Q = C′V, consequently

$$-\,C dV/dt = 4\pi\lambda_+ C'V, \quad . \quad . \quad . \quad (4)$$

a relation from which λ_+ can be determined. A similar experiment with the rod positively charged yields λ_-. The use made of the values found for the polar conductivities and an adaptation of the Gerdien method due to Schering are described in Chapter III, § 30.

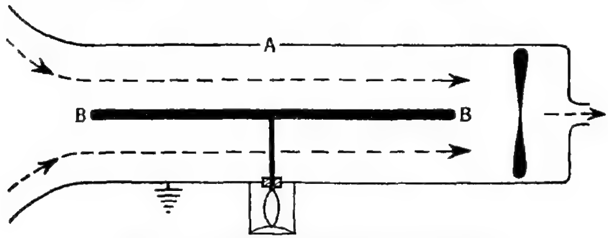

Fig. 2.—Gerdien conductivity apparatus.

§ 4. The Ionising Agencies

Three principal agencies are responsible for the ionisation of the lower atmosphere: radiations from radioactive substances in the surface of the earth, radiations from radioactive matter present in the air itself, and the penetrating radiation. For a discussion of their relative importance it is convenient to use the symbol q to represent the number of "pairs of ions produced per c.c. per second in air at N.T.P.," a quantity which will be referred to as the ionising power of the radiation. This phrase in inverted commas is usually represented by the symbol I. Thus the statement $q = 4$ I means that, at the point in question, the radiations create four pairs of ions per c.c. per second, if the medium there is air at N.T.P. The contributions of the three different agencies just mentioned to the total ionising power, q, are distinguished by the symbols q_e, q_a, and q_h.

§ 5. The Effect of Radioactive Matter in the Earth

Uranium, thorium, and their products are widely distributed through the earth's crust, whose surface layers therefore emit α, β, and γ rays into the air. Since the α rays can only emerge from a very thin layer and can only ionise the first few cms. of the air, their effect is inconsiderable. The β rays can come from greater depths and penetrate much further into the atmosphere. Calculations based upon average values for the measured concentration of radioactive matter in the earth indicate that an effect ranging from 1 I at the surface to 0·1 I at about 10 metres above it is due to these rays. The γ rays are a more important factor, for they can come from still greater depths and consequently from a still larger quantity of radioactive material. It is estimated that they produce on the average 3 I at the surface, 1·5 I at 150 metres above it, and 0·3 I at a height of 1 kilometre. The recent discovery of the γ ray activity of potassium introduces another source of ionisation, for the potassium content of the earth's crust is considerable. No very detailed survey of this question has yet been made, but measurements indicate that at a height of 10 metres the contributions of radium, thorium, and potassium to the γ ray ionisation of the air are in the ratio of 1 : 1 : 0·7, in accord with sampling of the surface content of the earth.[5] The average value for the ionising power of the earth-radiations near the ground has been found directly to lie between 2 and 10 I, which is in accord with the measurements of the radioactive content of the earth's crust. Local surface and geological conditions naturally have a marked effect upon the value of q_e at any given spot.

The radioactive content of sea-water is found to be negligible in comparison with that of soil and rocks, and over the oceans the earth-radiation effect is very small indeed.

§ 6. The Effect of Radioactive Matter in the Air

The atmosphere contains a considerable quantity of the radioactive emanations, radon and thoron, and their successive products. A wire charged negatively to a potential of a few hundred volts and exposed to the air collects an easily detectable quantity of radium A and thorium A by recoil from the disintegration of these emanations. The gases themselves arise from the decay of radium and thorium within the earth, from which they escape by diffusion, by thermal convection, and as a result of decreases in the external atmospheric pressure. The greater portion of the ionisation thus produced is due to the α rays from the emanations themselves and from their A and C products, for here, unlike the case of the radiations from the earth, there is no absorbing layer between the disintegrating atom and the air. The radioactive gases and their products are fairly evenly distributed by atmospheric turbulence through the lower air. This "air-radiation" effect can be calculated from radioactive data if the emanation content of the air is determined. The emanations may be removed from a measured volume of air by drawing it through tubes cooled with liquid air or filled with coconut charcoal. The quantity of emanation condensed or absorbed is then determined by the effect produced in an ionisation electroscope, corrected for the decay of the material during the time of the experiment. A more direct method is to fill a closed ionisation chamber first with ordinary air and then with air which has been freed by the above method from the emanations and from their products, and to determine the saturation currents in the two cases. If these are i_1 and i_2 respectively, v the volume of the vessel and e the elementary charge, the value of the air-radiation ionising power, q_a, is given by $i_1 - i_2 = q_a ve$.

The results of such measurements agree with estimates based upon observations of the average emanation

content of the air in placing the air-radiation effect at about 5 I in the neighbourhood of the earth's surface. Of this about 60 per cent. is due to radon and its α ray products and the remainder to the thoron series. It varies somewhat with locality and with those meteorological conditions which determine the rate of escape of the emanations Apart from the effect of strong upward winds in elevating the richer surface layers, the emanation content of the air should decrease fairly rapidly with height. Direct information on this point is difficult to obtain, but what exists indicates that the air-radiation effect falls to less than 2 per cent. of its surface value at a height of 5 kilometres.

The effect, like that of the earth-radiation, is extremely small over the oceans, where the emanation content of the air is only about 1 per cent. of its value over land areas.

§ 7. The Effect of the Penetrating Radiation

The third factor in the ionisation of the atmosphere is a radiation of peculiar interest, whose origin and nature are still the subject of much investigation. It has a penetrating power which is considerably greater than that of the most penetrating γ rays from radioactive bodies, and it travels to the earth's surface in a downward direction, undergoing a certain amount of absorption on the way. Its ionising power at sea-level over both land and ocean areas lies between 1·5 and 2·5 I. Though these figures correctly suggest that the absolute value of the effect is not accurately known, relative observations can be made with great certainty and show that the ionising power of the radiation at a given height above sea-level is practically independent of geographical position and alters only slightly with meteorological conditions such as pressure. The ionisation due to this radiation increases rapidly with height above sea-level, rising ten-fold in the first 5 kilometres. It is discussed in greater detail in the next chapter.

§ 8. The Ionisation Inside a Closed Vessel

It is a matter of considerable difficulty to determine separately the ionising powers of the earth-radiation and of the penetrating rays. Calculations of the former must be based upon measurements of the radioactive content of the earth's crust, and these are necessarily tedious and rough. To proceed further, one must examine the ionisation produced in the air inside a sealed metal chamber under conditions which are modified from those existing in the free air by the presence of the metal walls. This involves correction to free-air conditions. A further disadvantage is that a new ionising agency is introduced, the radioactive matter in and on the inside walls. The ionising power of this "wall-effect," or the "residual ionisation" of the vessel as it is called, is a constant depending on the material of which the walls are made and the cleanliness of the inner surface. It is usually represented by the symbol q_0.

The total ionising power of all the radiations operating upon the gas in a closed vessel may be called q' and the modified effects of the three main ionising agencies q'_e, q'_a, and q'_h : we then have

$$q' = q'_e + q'_a + q'_h + q_0. \qquad (5)$$

We may suppose the vessel to take the form of a sealed ionisation chamber filled with air at N.T.P. which has been freed from radioactive matter. The saturation current carried by the ions to the central collecting electrode is then given by $i = q've$, where v is the volume of the vessel and e the elementary charge on an ion.

Various methods have been employed to separate out the four terms on the right-hand side of equation (5) from one another. The most satisfactory of these begins by determining q_0. The instrument, whose form is described in the following chapter, is taken to a large snow-fed lake whose radioactive content, like that of the sea, is negligible. When it is sunk to the depth of a few feet in such a lake, the effects of earth and air-

radiations are non-existent and equation (5) becomes $q' = q'_h + q_0$; on lowering the vessel still further, q'_h diminishes as the penetrating radiation becomes absorbed in the increasing thickness of water, and it is possible to reach depths at which the saturation current is practically constant and unaffected by further sinking of the instrument. At this stage $q' = q_0$, and the residual ionisation is determined. By combining this observation with that made close to the surface of the lake, q'_h may be found.

Since the effects of the penetrating radiation and of the walls of the vessel are both independent of local conditions other than height above sea-level, they remain the same when the instrument is transported to a site on land at the same level. A measurement of the saturation current in the chamber under land conditions gives q' as stated by equation (5), and hence the sum of the earth and air effects, q'_e and q'_a can be found In actual practice the air-radiation effect within a closed vessel is extremely small: the walls exclude α and β rays coming from outside and reduce q'_a to the ionising power of the γ rays emitted by the B and C products of the radioactive emanations. This amounts to but 4 per cent. of the effect produced by the air-radiation in the free atmosphere, in the case of a vessel with steel walls a few millimetres thick. It is therefore sufficient to determine the air-radiation effect, q'_a, by calculation based upon a knowledge of the emanation content of the air and of the nature of the walls of the vessel. The fourth factor, the effect of the earth-radiation, q'_e, is then obtained by substitution in equation (5).

To pass from these closed vessel determinations of the effect of the radioactive matter in the earth to the effect to be expected in the free air, experiments may be made with screens of different thicknesses placed around the ionisation chamber and the values thus obtained for the earth-radiation effect extrapolated to zero thickness. In doing so account has to be taken of the effect of the screens upon q'_h.

Until more is known of the properties of the penetrating radiation, the relation of its ionising powers inside and outside a closed vessel cannot be stated with any certainty, but it is probable that unless the walls are very thick they will differ little from one another.

The greater part of the residual ionisation arises from α rays and takes place close to the walls; q_0 is therefore the effect of the residual ionisation near the walls divided by the volume of the vessel.

§ 9. Summary of Ionising Powers of Various Agencies

The table on opposite page gives a rough idea of the effects of the three main factors in the ionisation of the air at sea-level, both in closed vessels of steel, a few millimetres thick, and in the free atmosphere. The values given refer to measurements at about a metre from the surface of the earth; they are averages and, except in the case of the penetrating radiation, are subject to considerable fluctuations with local conditions.

It is noteworthy that the rate of production of ions over land areas is about six times as great as over the oceans, where the ionisation is practically all due to the action of the penetrating rays. Over land close to the surface of the earth, these rays are much less important, but at greater heights the effects of the earth and the air-radiations fall off rapidly as we have seen, so that here also, from a height of one of two kilometres upwards, the penetrating radiation is the chief agent in making the air a conductor.

§ 10. The Rate of Disappearance of the Small Ions

The number of ions produced by the agencies we have discussed will not increase indefinitely as time goes on. Even in pure filtered air, free from nuclei, the small ions disappear by diffusion to the walls and by recombination with one another. Experiments have shown

TABLE I

IONISING POWERS AT THE EARTH'S SURFACE, SEA-LEVEL

Ionising Agent.	Ionising Power in Ion-pairs per c.c. per sec. in Air at N.T.P.			
	Unscreened Closed Vessel.		Free Air.	
	Over Land.	Over Sea.	Over Land.	Over Sea.
Radioactive matter in the earth . .	4·0	0	4·0	0
Radioactive matter in the air . . .	0·2	0	5·0	0
Penetrating Radiation	2·0	2·0	2·0	2·0
	Total ionising power for the free air .		11·0	2·0

that when the diffusion loss can be neglected, the equation $dn/dt = q - \alpha n^2$ applies to such filtered air, n being as before the number of positive or negative small ions per c.c., q the number of pairs of ions produced per c.c. per second, and α the so-called recombination constant of small ions. In the course of time a state of equilibrium will be reached, in which the processes of creation and disappearance of ions balance one another; then $dn/dt = 0$ and the equilibrium equation $q = \alpha n^2$ holds. Applying this equation to the free air, taking $q = 11$ I from the preceding section and $\alpha = 1{\cdot}6 \times 10^{-6}$ cm.^{3}/sec. from laboratory experiments, we find $n = 2622$. This is very much larger than the number of small ions actually observed, which is round about 600 per c.c. The discrepancy is due to neglect of the important part

played by the Aitken or condensation nuclei in capturing the small ions.

An equation applicable to air containing nuclei was first given by von Schweidler [6] in the form

$$dn/dt = q - \alpha n^2 - \gamma n Z,$$

where γ is a constant and Z the total number of nuclei, charged and uncharged, per c.c. of the air. For equilibrium, when dn/dt is zero, it becomes

$$q = \alpha n^2 + \gamma n Z = n(\alpha n + \gamma Z) = n(\alpha n + \beta) \quad (6)$$

or

$$q = \beta' n \quad . \quad . \quad . \quad . \quad (7)$$

The quantity β' is known as the disappearance constant of the small ions. Experiment shows that in ordinary air, where nuclei are plentiful, the term αn is small compared with β; even over the oceans it is probably only one-seventh of β. Thus β', while not a true constant, for it depends upon the concentration Z of the nuclei in the air, is to a large extent independent of the number of small ions per c.c. Equation (7) thus states a linear law of recombination as opposed to the quadratic law which holds for filtered air.

If the action of the ionising agents were removed, the small ions would disappear at a rate $dn/dt = \beta' n$, giving $n = n_0 e^{-\beta' t}$, for the number present at any subsequent time. It is assumed in the integration that there are sufficient nuclei present to make β' independent of n. Von Schweidler has pointed out that the quantity $\theta = 1/\beta'$ can, by analogy with the equation for radioactive decay, be termed the average life of a small ion. Equation (7) then becomes

$$q\theta = n \quad . \quad . \quad . \quad . \quad (8)$$

in which θ should vary inversely as the number of nuclei per c.c. of the air.

In a development of this theory put forward by J. J. Nolan and his co-workers, [2, 1] separate consideration is given to the capture of small ions by charged nuclei of opposite sign, leading to neutralisation, and by un-

charged nuclei, leading to the formation of large ions. For the small positive ions, they write

$$dn_+/dt = q - \alpha n_+ n_- - \eta_1 n_+ N_0 - \eta_2 n_+ N_- = 0, \text{ for equilibrium} \quad . \quad (9)$$

Here N_0 and N_- are the numbers of uncharged and negatively charged nuclei per c.c. respectively, while η_1 and η_2 are the recombination coefficients for the two types of capture. By a theoretical discussion, too long to be reproduced, this equation is converted into the simpler form $q = [\eta_2 Z/2{\cdot}1]n$, which corresponds to Schweidler's equation (6) and again indicates that the average life of an ion should vary as $1/Z$.

This average life θ is found over land areas to lie between 10 and 60 seconds according to the purity of the air, the higher values being associated with observations made in the country. Over the oceans it may be as much as 230 seconds, according to the measurements of Hess [7] and of Mathias [8] on the island of Heligoland. The table below, taken from Hess, illustrates the great difference in the equilibrium between small and large ions brought about when the wind direction changes; the first set of observations was made when the wind blew from the nuclei-laden European mainland, the second when it blew from the polar regions:—

TABLE II

Observations of the State of Atmospheric Ionisation (Hess, Heligoland)

Wind.	No of Small Ions per c.c (n).		No. of Large Ions per c.c.(N).		Total No. of Nuclei.	Average Life of Small Ion, secs.	
	n_+	n_-	N_+	N_-	$Z = N_0 + N_+ + N_-$	θ_+	θ_-
S.W. (Land)	220	267	1480	1510	6300	22	22
N.N.W. (Arctic)	794	843	200	200	1100	115	204

These authors find that for values of θ lying between 36 and 100 seconds, Schweidler's linear equation (8) is valid; in purer air, where θ is greater than 100 seconds, the simple equation no longer holds, for the quadratic recombination term αn^2 then becomes significant.

Experiments by P. and J. Nolan [9] and their co-workers have, however, suggested that the form of the accepted equilibrium equation is incorrect and should be replaced by a new empirical relation $q = n(\alpha n + \zeta\sqrt{Z})$, where ζ is a constant. For air with a sufficient concentration of nuclei this reduces to $q = [\zeta\sqrt{Z}]n$, which would imply that the mean life of an ion varies as $1/\sqrt{Z}$. It is therefore possible that the processes governing the disappearance of the small ions in the atmosphere are more complex than has hitherto been supposed. Account may have to be taken of variations in the actual nature of the nuclei, of complexities in the small and the large ion families, such as are indicated by the existence of "intermediate" ions, of the possible occurrence of multiply-charged large ions when the number of nuclei is small,[10] and of the effect of actual dust-nuclei.

§ 11. Methods of Determining the Average Life of a Small Ion

The average life, θ, of the small ions can be determined for a sample of air enclosed in an ionisation chamber by two methods suggested by Schweidler.[6] The first involves the separate measurement of the ionising power, q, of the ionising radiations, and of the number of small ions per c.c., n, and substitution in the equation $q\theta = n$. q is obtained from the saturation current, i, using the relation $i = qev$, where v is the volume of the vessel. The corrections to be applied have already been discussed (§ 8). The quantity n is found by allowing the air to stand undisturbed for a sufficient time for equilibrium between small ions and nuclei to be attained and then applying a strong field for a few seconds to sweep the small ions on to the collec-

ing electrode. The second method involves the measurement of the currents flowing in an ionisation vessel for various values of the applied potential difference.[11]

§ 12. The Ionisation-balance in the Lower Atmosphere

It is of considerable importance to decide to what extent the various factors in the creation and destruction of ions account for the state of ionisation actually observed in the air over land and sea. Any failure in this respect would suggest that new processes must be sought.

The evidence available indicates, however, that the chief factors in the ionisation-balance have been discovered. Over land areas where nuclei are usually plentiful, the Schweidler equilibrium equation, $q\theta = n$, applies. An average value of 60 seconds for θ, the mean life of a small ion, and of 650 per c.c. for the small ion concentration, requires that the rate of creation, q, should be 650/60 or 10·8 ions/c.c./sec. in satisfactory agreement with direct measurements of q, shown in Table I. No direct measurements of θ have yet been made over the oceans. The concentration of nuclei in mid-ocean has been estimated by Hess, from measurements on air from the Arctic, to be such as to give an average life of 230 seconds to a small ion. Since n was found on the *Carnegie* to amount to about 550 small ions/c.c., the ionising power at sea should amount to 550/230 or 2·4 I. The agreement of this figure with the value 2·0 I shown in Table I. is reasonable, in view of the roughness of the estimate of θ used, and of the present uncertainty in the value of q itself.

Hess's estimate of the concentration of Aitken nuclei over the oceans, about 800 per c.c., is borne out by direct measurements on the *Carnegie*, which give means of 870 and 774 per c.c. in the Atlantic and P...ic measure-
respectively, and by a recent count ...eaks, or in balloons,
the Atlantic. ...ust and portable, as

The problem of the ionisation

cannot be said to be completely solved until agreement is reached on the form of the interaction-relation between small ions and nuclei. The general nature of the factors in the ionisation-balance has, however, been established, and an explanation can be given of the paradox that the conductivity of the air undergoes no large change as one passes from mid-ocean to land, despite a six-fold increase in the rate at which ions are formed. The increased birth-rate of the small ions is counterbalanced by an increased death-rate, owing to the greater concentration of predatory condensation-nuclei over the land areas.

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[3] Boylan, *ibid.*, A, 37, p. 58 (1926).

[4] C. T. R. Wilson, *Dict. of Appd. Physics*, Art. *Atmospheric Electricity*, vol. 3, p. 84.

[5] Suckstorff, *Naturwissenschaften*, 19, p. 87 (1931).

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[7] Hess, *Gerland's Beit. z. Geophysik*, Bd. 22, p. 256 (1929).

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CHAPTER II

THE PENETRATING RADIATION

UNTIL the year 1911 it was generally supposed that the ionisation inside a closed vessel was entirely due to radiations from the walls, and from radioactive matter in the earth and the outside air. It was therefore to be expected that this ionisation would decrease when the vessel was raised above the surface of the earth. This was tested in balloon ascents by Hess in 1911, and Kolhörster in 1913, who found that the diminution with height was only appreciable during the first kilometre, after which an increase occurred. These observations led Hess to postulate an ionising radiation coming from above and suffering partial absorption in the atmosphere.

Something has already been said of this radiation in § 7; the accurate determination of its properties has proved a difficult matter. Its great penetrating power implies that the ionisation it creates within a closed vessel is so small as to require specially refined technique to measure accurately in the presence of the other ionising agencies discussed in Chapter I. Further, the fact that it comes from above and suffers some absorption necessitates investigations at high altitudes so that any softer and more easily absorbed components which do not reach sea-level may be examined. Such measurements must be made on mountain peaks, or in balloons, under conditions which require robust and portable, as well as sensitive, instruments.

§ 13. Experimental Methods of Measuring the Intensity of the Radiation

The most extensive measurements of the intensity of the radiation have been made by the ionisation method. The first point to be attended to is to eliminate as far as possible the disturbing effects of the earth and air-radiations. This can best be done by sinking the ionisation chamber beneath the waters of a lake, or by surrounding it with a thick shield of non-radioactive iron or lead and correcting for the small fraction of the γ rays which are still able to enter the chamber.

The ionising power of the penetrating rays at sea-level is about 2·0 I (ions per c.c. per sec.), so that their contribution to the saturation current in a vessel of volume 1 litre, filled with air at N.T.P., would be $3{\cdot}2 \times 10^{-16}$ amperes. In order to increase this effect, it has become the general practice to fill the vessel with carbon dioxide at a pressure of 30 atmospheres or more.* This carries with it an important advantage, in that it considerably reduces the effect of the residual ionisation, which, we have already seen, is caused by α rays from the walls. The range of these rays in CO_2 at high pressure is small, and they produce all their effect within a few millimetres of the walls themselves. Here the electric field driving ions to the central collecting electrode is weakest and cannot separate out more than a small fraction of those produced before they recombine. A further decrease in the residual effect can be obtained by using a wire cage as an electrostatic screen to form the walls of an inner ionisation chamber. This cuts down the amount of wall material productive of undesirable α rays (Hoffmann, Fig. 4).

The quantity actually measured in these experiments is the sum of the penetrating ray and the residual

* Increasing the pressure does not give a proportional increase in ionisation. For 30 atmospheres the effect is only 14 times that at N.T.P. and the actual factor is somewhat uncertain. Hence the difficulty of finding the absolute value of q_h.

ionisations, q_h and q_0, and fluctuations in the latter limit the accuracy with which q_h can be found. These arise from the fact that in a given time-interval the number of α particles emitted from the walls undergoes variations about a mean value which are subject to well-known statistical laws. To smooth them out it is generally necessary to make observations over a period of several hours. For accurate measurements in a shorter time instruments have been devised by Hoffmann [1] and by the author [2] which are sensitive enough to show the sudden increases in the ionisation due to single

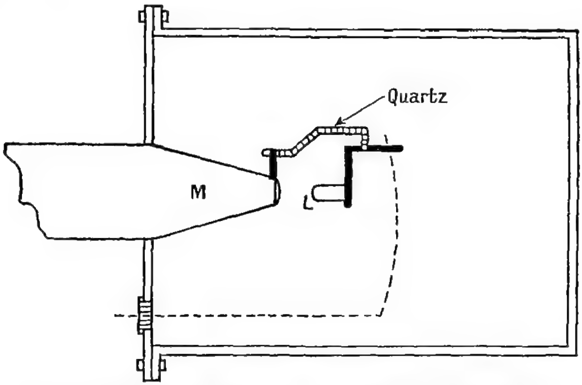

FIG. 3.—Kolhörster ionisation chamber and electroscope.

α particles and so to determine the residual ionisation directly.

The portable instrument developed by Kolhörster is shown diagrammatically in Fig. 3. The ionisation chamber is of steel, coated inside with zinc, and is filled with purified air at atmospheric pressure. Two small loops L of platinised quartz fibre, mounted upon a rod held in a quartz insulator, form the central electrode and can be charged by a device operated from outside by an electromagnet (shown dotted). Their potential is determined from their distance apart as read in the

eye-piece of a microscope M. In Millikan's instrument quartz fibres are also used as the central electrode, but the gas-filling is CO_2 at a pressure of 30 atmospheres. In both cases the quartz and invar-steel framework supporting the fibres is designed to reduce the effects of temperature variations. The chief disadvantage of this type of chamber is the small volume permissible in

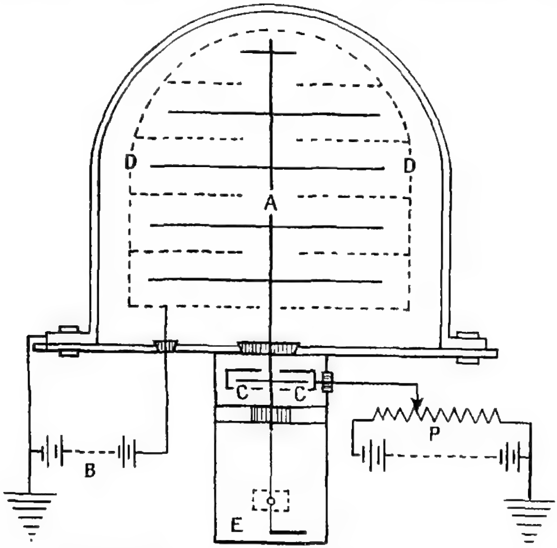

FIG. 4.—Hoffmann's null method for the measurement of the ionisation due to the penetrating radiation.

view of the pressure and the use of a high-power microscope. The penetrating radiation ionises the gas through the agency of fast-moving particles (§ 17) which fluctuate in number like the α particles from the walls. To smooth them out in the case of small ionisation vessels necessitates observations extending over many hours.

Another type of apparatus, due to Hoffmann,[3] employs

an electrometer which is outside the ionisation vessel (Fig. 4). The chamber is filled with CO_2 at a pressure of about 30 atmospheres, the residual radiation is reduced by the use of a wire cage for the walls of the ionisation vessel proper, and saturation conditions are obtained by adopting the branched form of central electrode A and cage D. The electrometer E is of a specially sensitive type developed from the quadrant electrometer. It serves as a null instrument, for arrangements are provided to maintain the whole central system at zero potential. The cage is kept permanently charged by means of the battery B, and the charge carried to the central electrode by the ions is compensated for by an equal and opposite charge induced upon the inner plates of the small condenser C by means of the potentiometer P. The amount of charge collected during a given time can be found from the capacity of C and the potential applied to the outer plates. Both C and the electrometer are kept evacuated. Though more complicated and less portable than the first type, this instrument has the advantages of freedom from all insulation losses on the collecting system and of greater volume.

The more recent portable instruments show a reduction in the ionising power of the residual wall radiation; in Millikan's 1926 form the ratio of this quantity to the ionising power of the penetrating radiation at sea-level was 4·5/1; in the form used by him in 1930 this had been reduced to 1/30. The ratio for Hoffmann's instrument is less than 1/5.

§ 14. The Ionisation-depth Curve in Air and Water

In studying the variation of the ionisation intensity of the radiation above and below sea-level, it has been found convenient to measure distance downwards from the top of the atmosphere. This distance is usually expressed in equivalent metres of water, i.e. in terms of that column of water which would possess the same

mass per unit area of cross-section as the column of air above the point of observation. Thus sea-level is 10·33 equivalent metres of water below the top of the atmosphere.

Observations of the ionisation at various depths below the top of the atmosphere are most conveniently made by sinking an ionisation vessel in a lake situated at a very great altitude. The results when plotted against the total depth, equivalent and real, of water below the top of the atmosphere constitute an ionisation-depth curve. An extensive series of investigations has been carried out by Millikan and Cameron to determine the shape of this curve, using lakes on mountain peaks in both North and South America.[4] Thus they were able to examine the absorption of the rays from near the surface of Gem Lake, California, where the depth below the top of the atmosphere was 7·4 equivalent metres of water (altitude 3000 metres), down to 81 actual metres below the surface of the lake, 86 equivalent metres from the top of the atmosphere. Owing to the need for some water-screening to cut down the air-radiation effect, q'_a, trustworthy observations began at a total depth of some 8 metres.

They have repeated these observations in other lakes at lower altitudes and have been able to show that at the same total depth below the top of the atmosphere the ionisation is always the same. This indicates that within the reasonably high accuracy of their measurements, the absorbing power of the additional thickness of air involved in the observations at lower altitudes is in fact that of the equivalent layer of water. As a consequence of this result measurements made at still greater altitudes on mountain peaks and in balloons can be plotted on the ionisation-depth curve for water with some degree of confidence that the absorption in the air can be reduced to that in water in this way. Observations made in air alone, however, require the instrument to be shielded with lead from the external earth and air-radiations and have not only to be cor-

rected for the small fraction of these which can still penetrate the shields but also for the absorption of the penetrating radiation in the lead. For these reasons they are not so reliable for plotting on the curve as the measurements made in water.

The ionisation-depth curves obtained by Millikan and Cameron are shown in Fig. 5, the portion below 30 metres being inset on a larger scale. The observations made in Lake Arrowhead (altitude 1700 metres) and Gem Lake (altitude 3000 metres) constitute the main "water" curve. The curve marked "land" shows the

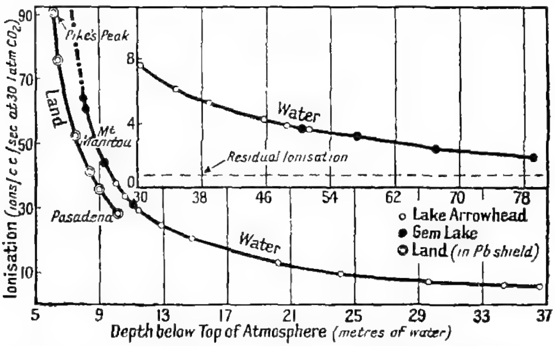

Fig. 5.—Ionisation-depth curve in air and water (Millikan and Cameron).

land observations made at various heights from sea-level to the top of Pike's Peak, California (altitude 4700 metres, 6·2 equivalent metres of water below the top of the atmosphere), inside a thick lead shield. By comparing the abscissæ of the land and water curves for the same ordinate, the water-equivalent of the shield at the higher altitudes was found to be 1·22 metres. The dotted portion of the water curve is obtained by adding this amount to the equivalent depth of the land observations.

Even at a total depth of 80 metres the water curve

is not parallel to the horizontal axis and the value of the residual ionisation (1·2 ions/c.c./sec. in CO_2 at 30 atmospheres) has to be obtained by extrapolation.

These measurements show clearly the very great penetrating power of the radiation, for the most penetrating known gamma rays (from thorium C″) would be reduced to one-fiftieth of their original intensity in their

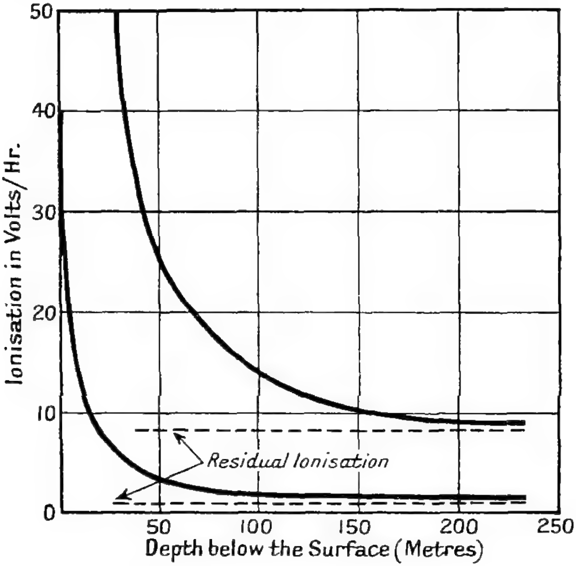

FIG. 6.—Ionisation-depth curve (Regener).

passage through a single metre of water. The lower part of the ionisation-depth curve has been extended still further by Regener [5] in experiments on Lake Constance. His ionisation vessel had a volume of 34 litres and was filled with CO_2 at a pressure of 30 atmospheres. This was surrounded by a protecting tank filled with water from the surface of the lake which formed a screen

1 metre in thickness and prevented possible variations with depth in the small radioactive content of the surrounding water from affecting the measurements. With this arrangement Regener has followed the radiation down to a depth of 230 metres of water below sea-level, the large volume of the chamber enabling accurate measurements to be made even at great depths. At 230 metres the curve was found to be still falling and could still be determined with an error of the order of 2 per cent. The movement of the fibres of the electrometer was recorded automatically by photographing their position at intervals of one hour. Regener's curve is shown in Fig. 6, with the ordinates of the portion below 30 metres repeated on a tenfold scale.

§ 15. Analysis of the Ionisation-depth Curve

The form of the ionisation-depth curve suggests that it may be made up of three or four separate curves corresponding to component radiations of different penetrating powers. On this view the initial steep portion would correspond in the main to an easily absorbed component which does not reach sea-level in any quantity compared to the others, while only the most penetrating component can reach to the great depths examined by Millikan and Regener. In endeavouring to analyse the curve in this way it has been assumed that the process of absorption of a parallel beam of the s-th component is represented by the well-known equation $-dI = \mu_s I dH$, where $-dI$ is the decrease in the ionisation I due to passage through a thickness dH of the absorbing medium and μ_s is a constant, characteristic of the radiation and of the medium, which is called the absorption coefficient of the s-th component. Integration of this equation gives $I = I_0 e^{-\mu_s H}$. The significance to be attached to μ_s will be discussed later.

There is, however, no reason to believe that the radiation strikes the top of the atmosphere as a parallel

beam, and we must assume incidence with equal intensity in all directions and take account of the increased thickness of air or water to be traversed as the angle of incidence is increased. When this is done, it is found that the relation between the ionisation I at a depth H and that which would be found at the top of the atmosphere, I_0, takes the form $I = I_0 \int_1^\infty [e^{-\mu_s H x} . dx/x^2]$.* Tables of this integral have been given by Gold.

The analysis of the ionisation-depth curve must be carried out by a trial and error method, for not only the number of components but the absorption coefficient μ_s and the initial intensity I_0 of each has to be chosen to fit the curve. Analyses of this kind have been made by Millikan and Cameron, and by Regener, whose results are shown in the following table :—

TABLE III

ANALYSES OF THE IONISATION-DEPTH CURVE INTO COMPONENTS

Compt.	Millikan.		Regener.	
	Absorption Coefficient, μ_s.	Intensity at Top of Atmosphere, I_0.	Absorption Coefficient, μ_s.	Intensity at Top of Atmosphere, I_0.
1	0·80 metres^{-1}	141,000	?	?
2	0·20 ,,	130	0·21	130
3	0·10 ,,	80	0·073	51
4	0·03 ,,	33	0·02	6·5

The values of μ have been expressed in reciprocal metres of water ; in the same units the mean absorption coefficient of the gamma rays from radium B + C

* See Rutherford, *Radioactive Substances and their Radiations*, p. 262, where $x = \sec \theta$.

after filtration through 10 mm. of lead would be 5·4. Since Regener's experiments did not extend above sea-level he has not been able to determine the constants of the most easily absorbed component.

The agreement obtained by these independent observers for the other three components is striking; slight discrepancies in μ may be explained by the uncertainty in the exact value of the residual ionisation, for the curves are still falling at the lowest depths reached. The difference in relative intensities may be due to some selective action arising from the difference in the size of the chambers used.

§ 16. Ionisation-depth Curves in Heavy Metals: the Transition Zone

The absorption of the radiation in heavy metals has been studied by placing metal screens round the ionisation vessel. Since this corresponds to sinking it in a lake of the metal, the results obtained should agree with the ionisation-depth curve taken in water when the thickness of each screen is expressed in equivalent metres of water. It is, however, found that the absorbing power of a metal such as lead is initially very high. The curve falls abnormally rapidly at first (for 10 cm. in the case of Pb); afterwards it runs parallel to, but lower than, the water curve, additional screens producing the same reduction as their water-equivalents.* It has been shown that this effect is due to the transition from air to lead, and something similar is exhibited in all transitions from one medium to another. Fig. 7, after Steinke,[6] shows the ionisation-depth curves of aluminium and lead as thick lines. The effects of adding an inside layer of Pb after 28 cm. of Al have been used, and of adding inside layers of Al after 4 and 10 cm. of Pb, are

* In this work the water-equivalent is that thickness of water with the same number of extra-nuclear electrons in a column one sq. cm. in cross-section. Millikan's water-equivalents (§ 14) are mass-equivalents.

shown by the dotted lines. The increase in the ionisation in two of these cases is very remarkable. It is suggested that the transition zone effect is due to an alteration in the equilibrium between the primary rays and their secondary electrons (§ 20), but no quantitative explanation can yet be given. The phenomenon is of great importance in connection with the problem of the nature of the radiation.

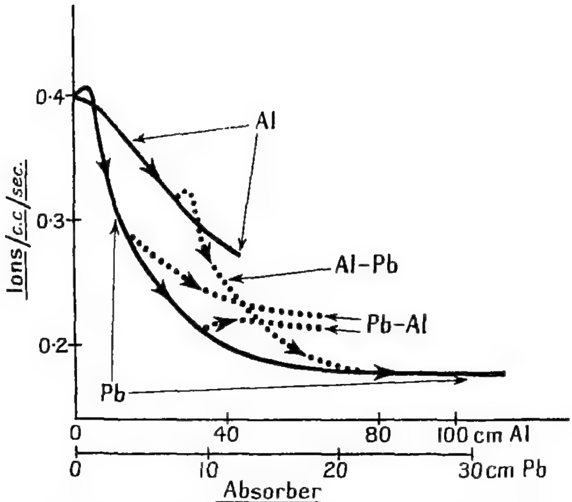

Fig. 7.—Transition-zone effects in Pb and Al (Steinke).

A striking instance of the initial drop in the absorption curve in lead is afforded by a screen used by Millikan in mountain observations. Its calculated water-equivalent was 85 cm., but a comparison with the ionisation-depth curve in water showed that it was effectively 170 cm. at sea-level and 122 cm. at a height of 5000 metres.

§ 17. Photographic Registration and Counting of the Ionising Particles

The ions produced by the rays can be studied by means of the Wilson cloud chamber, in which a water-drop is condensed on each. Skobelzyn [7] has shown in this way that much of the ionisation is due to the action of fast-moving particles which travel in a downward direction at the rate of about one particle per square centimetre per minute at sea-level. The photographs of their tracks show that they create some 40 ion-pairs per centimetre of their path in ordinary air.

It is not yet certain whether all the ionisation arises

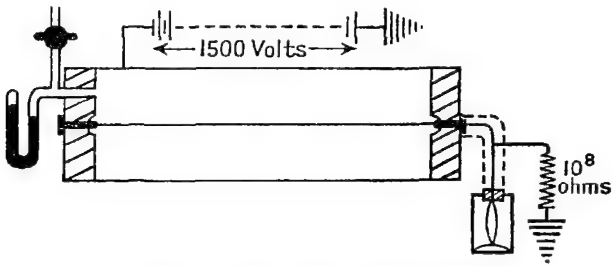

Fig. 8.—Geiger-Müller tube-counter.

from particles of this kind. A number of the tracks show a paired relation which may be significant, two or more particles entering the chamber during one expansion in directions making a small angle with one another.

The passage of these ionising particles can be continuously recorded with the tube-counter of Geiger and Müller. In this (Fig. 8) the ionisation vessel is a tube of zinc and the collecting electrode a thin wire along the axis, coated with a semi-insulating layer of oxide or varnish. Conditions of pressure and potential difference are adjusted below the sparking value so that the entry of each ionising particle gives rise to a sudden

pulse of current of very short duration. The advantage of this device is that the effect of an α particle is equal to that of a β particle, instead of 200 times greater as in the ordinary ionisation chamber. The same cleaning and screening precautions as have been described for the latter must be taken. This counter has been

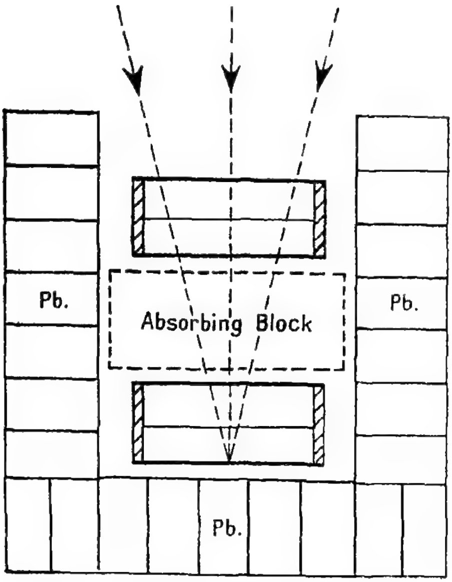

FIG. 9.—The Bothe-Kolhörster experiment.

used by Regener [5] to confirm the readings on his ionisation-depth curve; the submerged counter was joined to an arrangement which recorded each pulse automatically.

An apparatus which counts the penetrating ray particles only has been developed by Kolhörster and Bothe (Fig. 9). Two counters are mounted vertically

above one another, and account is taken only of those pulses which coincide in time with one another and are considered to be due to the passage of particles which are able to penetrate both counters. The coincidences were originally determined by recording the sudden deflections of both electroscopes on the same moving photographic film. In a development due to Bothe,[8] the two counters are joined to the two grids of a screened-grid valve which is so arranged that a pulse of plate current occurs only when both grids are practically simultaneously changed in potential.

§ 18. The Bothe-Kolhörster Experiment

Bothe and Kolhörster[9] have used this arrangement to determine the absorption coefficient of the ionising particles themselves. A block of gold 4·1 cm. thick was placed between the counters, and the reduction in the number of coincidences was observed. The value found for the absorption coefficient was so close to that given by the usual ionisation chamber method as to lead them to conclude provisionally that the ionising particles are the primary radiation and not of secondary origin. In an extension of this experiment carried out by Rossi,[10] however, evidence was found of the actual production of fresh ionising particles when the radiation was absorbed. These presumably arise from a primary radiation of γ ray type.

§ 19. The Energy of the Ionising Particles

If the ionising particles are electrically charged, the deformation of their paths in a transverse magnetic field should give information as to their nature and energy. Several unsuccessful experiments of this kind have been interpreted as indicating that their momentum is too great for them to be bent in the fields used. Mott-Smith[11] in this way has placed a lower

limit of 2×10^9 *e*-volts * to their energy if they are electrons, and 10^9 *e*-volts if they are protons.

The Bothe-Kolhörster-Rossi experiment suggests that the effective range of the particles is about 10 equivalent metres of water or $7{\cdot}7 \times 10^5$ cm. of air at N.T.P. They produce some 40 ion-pairs per centimetre of air and each ion-pair costs the particle about 30 *e*-volts. Hence the total initial energy of a particle must be of the order of $40 \times 30 \times 7{\cdot}7 \times 10^5$ or $9{\cdot}2 \times 10^8$ *e*-volts.

In experiments carried out under thunderstorms,[2] the author has found that the strong electric fields within and outside the cloud are able to stop a large proportion of these particles. He concludes that their energy must be less than 5×10^9 *e*-volts, if they are electrons or protons.

Such energies are more than a hundred times greater than those of the fastest particles emitted by radioactive bodies. If the particles are electrons they must be moving with a velocity only about 40 metres per second less than that of light and their inertial mass is 2000 times their rest-mass, slightly greater than the mass of a slow proton. At such energies electrons and protons may be expected to give approximately the same ionisation per unit path and the same absorption coefficient.

It has recently proved possible to obtain magnetic bending of the tracks in a Wilson cloud-chamber.[18] The paired particles are found to be of two types, positively and negatively charged, presumably protons and electrons. Their energies range from 10^7 to 10^9 *e*-volts.

§ 20. The Nature of the Penetrating Radiation

In view of the difficulties in the way of determining the nature of the fast ionising particles and of some

* Electron-volts, the energy expressed in terms of the difference of potential in volts through which an electron must travel to attain this energy.

uncertainty as to whether there are not other ionising rays of much shorter range, any discussion of the nature of the penetrating radiation itself is purely tentative. In this section two of the many important points which have been raised in this connection will be outlined.

The first concerns the effects to be expected if the primary radiation is itself constituted of fast-moving charged particles. In such a case the earth's magnetic field should produce a deflection of their paths leading to a concentration of the particles in the higher magnetic latitudes. Brucke has shown that for a fast electron to reach the earth's surface below magnetic latitude 50° its energy must exceed 5×10^9 volts, and for it to attain the equator it must have twice this energy, even if loss of energy in passage through the air be neglected. This question is being investigated at the present time; Millikan [12] has shown that the sea-level intensities in latitudes 59° and 34° are the same within the limits of accuracy of his measurements (1 per cent.). These and other observations speak against a corpuscular nature for the primary rays.

The second point concerns the view that the ionising particles are secondary electrons arising from the passage of primary ultra-gamma ray quanta through the atmosphere, and the utilisation of absorption coefficients derived from the ionisation-depth curve to find the frequencies of these quanta.[13] On this view an ionising particle is produced when a quantum of energy $h\nu$ makes a Compton collision with an electron in an atom. The electron moves forward with energy $h(\nu - \nu_1)$ and the quantum recoils with a lower "degraded" frequency ν_1. Consideration of the equations governing the Compton scattering process shows that for high-frequency quanta the collision in most cases leaves quantum and electron moving forward along paths only slightly inclined to one another, so that the radiation at any point within the atmosphere should consist of a mixture of primary and degraded quanta of various orders (the results of one, two, and more successive

collisions) together with the fast electrons arising from collisions of primary and degraded quanta.

Absorption measurements are made upon this mixed beam and yield an absorption coefficient $\bar{\mu}$; what relation this bears to the absorption coefficient of the pure primary radiation, μ, can be determined only by involved calculation. Since the degraded quanta have frequencies ν_1, ν_2, etc., which are less than the primary frequency ν, they are more easily absorbed than the primary quanta. Thus, as the radiation travels through the air, there will come a stage when as many degraded quanta of each kind disappear in each centimetre of path as are produced by Compton processes in the same centimetre. The parent beam is then in equilibrium with its *quantum* products in the same way as a long-lived parent radioactive body such as radium reaches equilibrium with its descendants. From this stage onwards, say after a thickness equivalent to D cm. of water, the constitution of the quantum part of the beam will remain unaltered, and this part will disappear at a rate set by the rate of disappearance of the primary quanta. Until this stage is reached $\bar{\mu}$ cannot equal μ.

Actual absorption measurements, however, are concerned with the ionising electrons and not with the quanta which give rise to them. Consider the first scattering process only, which we may assume for simplicity to give rise to electrons all of the same range, R cm. of water. Then the ionisation at a depth H cm. of water below the top of the atmosphere will be due to electrons from points lying between H and $H - R$ cm. from the top. Thus the actual measurements reflect the state of the beam some distance above the point of observation and can give only an apparent absorption coefficient $\bar{\mu}$ which equals μ, provided that $H - R > D$. If this condition is not fulfilled, the number of electrons brought to rest in 1 cm. is less than the number generated in the same distance and $\bar{\mu}$ will be less than μ.

The ionisation-depth curve has been analysed into

four component curves, distinguished by different values of $\bar{\mu}$. The significance of these apparent absorption coefficients obviously depends upon how far the criteria for equilibrium are satisfied in each case. Since according to Millikan, only the least penetrating component reaches equilibrium with its electrons and degraded quanta, the values of $\bar{\mu}$ found for the others will be less than μ. The deduction of their true absorption coefficients is at present impossible, for it involves a wide extrapolation of the laws governing the passage of γ and β rays through matter. Recent experiments have shown that these processes are not completely accounted for by existing theories. In particular, high-frequency quanta interact with atomic nuclei in a manner which is likely to be very important in the case of the penetrating radiation. The paired proton and electron tracks referred to in §§ 17 and 20 show that the rays frequently disintegrate atomic nuclei. This process is not taken into account in the γ ray absorption formulæ. If the radiation does consist of a mixed beam of ultra-gamma quanta and fast electrons, the observed absorption coefficients do not in general approximate to the true coefficients, and it is not possible to deduce the quantum frequencies except perhaps in the case of the softest component.

§ 21. The Origin of the Radiation

The intensity of the radiation is remarkably constant, and no certain variation with solar or sidereal time has been established. Corlin [14] and others have put forward evidence of a sidereal time effect which decreases with depth below the top of the atmosphere and amounts at sea-level to about 2 per cent. of the mean intensity. The effect only shows in averages of a long series of hourly observations, for alterations in the effective thickness of the atmosphere (not always faithfully represented by changes in the barometer) and other causes give rise to larger and irregular fluctuations.

Hoffmann, with the sensitive instrument described in § 13, can find no evidence for a sidereal time variation. A suggested solar effect also awaits confirmation; if it exists at all it is very small ($\frac{1}{3}$ per cent).[16]

It is most natural to look for the source of quanta of such high energy in the interiors of the hot stars, where high temperatures may lead to the conversion of matter into radiation. This possibility has, however, been rejected owing to the absence of appreciable solar and sidereal time effects.

Another argument against it is provided by the estimated total energy reaching the outer atmosphere in the form of penetrating radiation. This is more than one-tenth, probably one-fifth, of the total energy arriving from all the stars as light and heat. If the radiation originated in stellar interiors, very little of it could escape unaltered and most of it would be turned into heat. Unless, therefore, the radiation is generated only in the outermost layers of the stars, it is not possible for it to convey anything like such a large fraction of the total stellar radiation as one-fifth.

Millikan [4] has considered the possibility of the radiation arising from the formation of more complex elements out of hydrogen in interstellar space. The mass of a heavy atom being less, by Δm, than the separate masses of its constituent protons and electrons, energy of amount $\Delta m \times c^2$ ergs would be radiated out as a quantum whose frequency is given by $h\nu = \Delta m \times c^2$, where c is the velocity of light. He considers that the four main components of the radiation (§ 15) represent the formation of the four most abundant groups of elements, apart from hydrogen, viz. helium, "oxygen" (C, N, and O), "silicon" (Mg, Al, and Si), and "iron" (Fe, Ni, and Co). These elements make up 99 per cent. of all known matter, excluding hydrogen. The atomic weight determinations of Aston with the positive ray mass spectrograph give Δm in each case, and thus the quantum frequencies, ν, can be calculated. The corresponding absorption coefficients, μ, are calculated by

extrapolating the relation which holds approximately in the gamma-ray region. The theoretical coefficients for the four groups are ·796, ·241, ·142, and ·075 m^{-1} H_2O, respectively, and are to be compared with the four values of the apparent absorption coefficients, $\bar{\mu}$, from the ionisation-depth curve (Table III), ·80, ·20, ·10, and ·03. Millikan attributes the disagreement in the last three cases to a failure on the part of the more penetrating components to reach equilibrium with their secondary electrons. This, as we have seen (§ 20), would cause $\bar{\mu}$ to be less than μ, but the difficulty of making a quantitative check on this point and the impracticability of any extrapolation of the $\nu - \mu$ relation have delayed acceptance of the theory.

C. T. R. Wilson [15] has suggested that part at least of the radiation may be of terrestrial origin and come from thunderclouds. A β particle of radioactive origin emitted upwards in the strong field within the electrified cloud would, he has shown, rapidly increase in energy, for the gain of energy from the electric field would exceed the loss of energy by ionisation along its path. Since he and others consider the electric field within the cloud to be in general directed downwards, a thundercloud should spray upwards a stream of "runaway" electrons of energy up to 5×10^9 e-volts (see § 42). These would either be bent down in the earth's magnetic field to strike the earth at distant points or would be converted by direct hits with atomic nuclei into ultra-gamma quanta. Attempts to find direct evidence of this effect are at present being made.

A suggestion that the dissolution of matter into radiation may be responsible for the more penetrating components of the rays has recently been supported by the argument that the smallest absorption coefficient found by Regener (·0205 m^{-1} H_2O) corresponds on certain reasonable assumptions (Jeans [17]) to the value calculated for the radiation arising from the complete dissolution of a particle within the nucleus of a heavy atom (·020 m^{-1})

Some workers consider that apparent contradictions in the results of experiments on the radiation arise from the fact that it consists of fast-moving "neutrons," proton-electron combinations endowed with no resultant charge but having a magnetic moment.

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CHAPTER III

ELECTRIC FIELDS AND ELECTRIC CURRENTS IN THE ATMOSPHERE

§ 22. The Fine-weather Field

IT was first found in the year 1752 by Lemonnier that the air above the earth is the seat of a persistent electric field during fine weather. The direction of this field shows that the earth carries a negative charge and the upper layers of the air a positive one. The state of affairs close to a flat portion of the surface of the earth can be expressed in three ways: (*a*) the earth carries a charge of surface density σ per unit area; (*b*) there is a vertical field of strength $F = 4\pi\sigma$, just above the ground; or (*c*) between two horizontal planes close to the ground there is a difference of potential

$$dV = V_{h+dh} - V_h = - Fdh = - 4\pi\sigma dh,$$

where $h + dh$ and h are the heights of the planes.

The quantity dV/dh is called the potential gradient and is positive, since σ is found to be negative. For the same reason, the field F is directed downwards, and this downward direction is by convention adopted as the positive direction for electric fields in the atmosphere. In these definitions it is assumed that the portion of the ground considered is flat and far removed from projections, such as trees and buildings, which would disturb the distribution of charge and concentrate the lines of force at certain points.

The average value of the fine-weather potential gradient is about 100 volts per metre; the corresponding value for the average charge density is $2{\cdot}7 \times 10^{-4}$ electrostatic units per square centimetre or ·0009 coulomb per square kilometre. The total fine-weather charge on the earth is of the order of 500,000 coulombs. Measurements of the fine-weather field are made by determining either σ or dV/dh; they are always expressed in terms of the latter, in volts per metre.

§ 23. Direct Measurement of the Surface Density of the Earth's Charge

The charge on such a limited portion of the ground as can be isolated for the determination of σ is so small that a sensitive electrometer must be used to measure it. Fig. 10 illustrates the "Universal portable electrometer" devised by C. T. R. Wilson for this purpose. A flat circular plate P, called the test-plate, is surrounded by a guard-ring and mounted flush with the surface of the earth but insulated from it. This plate is joined to a gold-leaf electroscope G, whose leaf moves inside a positively charged case and is observed through a telemicroscope. C is a variable cylindrical condenser, called a compensator, with its inner plate joined to the gold-leaf system, and its outer plate maintained at a constant potential, — V.

To make a measurement, the instrument is initially shielded from the earth's field by placing over it the earthed metal cover Q, indicated by the dotted lines. The reading of the leaf corresponding to earth or zero potential is found by momentarily earthing the plate P and the gold-leaf system with the key K. This earthing key is then withdrawn, the cover removed and the plate exposed to the field, with the compensator capacity zero. A negative charge thus appears by induction upon P, an equal positive charge is set free upon the central system, and the leaf moves inwards. It can be restored to its zero position if a negative charge is

induced on the central system by increasing the capacity of the compensator to a value C′. The charge on the test-plate is then $-$ C′V and the surface density, σ, is $-$ C′V/A, where A is the area of the test-plate.

In its original form the apparatus is small and portable,

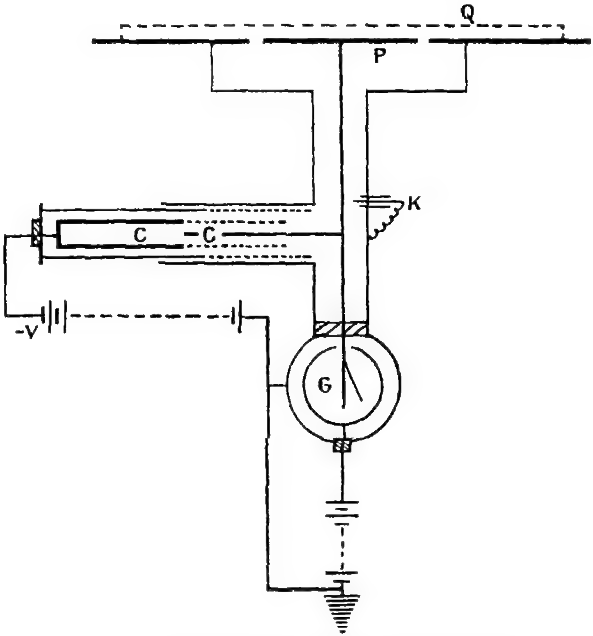

Fig. 10.—Universal portable electrometer (Wilson).

and the two batteries shown in the figure are replaced by small Leyden jars of silvered quartz, which keep their charge for a considerable time. For absolute measurements, with the plate mounted at ground level, a pit must be provided to house the observer ; very often relative measurements are made with the instrument

mounted on a tripod, and the factor necessary to reduce the observations to standard conditions determined from simultaneous observations in a pit or with a potential gradient method. Since the lines of force are concentrated upon any projection above the earth, the factor is greater than unity; in the case of a tripod 1 metre high it amounts to 3 or 4. A useful method of calibrating the instrument is to place it in artificial electric fields of known strength. Two large flat metal sheets, one flush with, and the other above, the test-plate, are charged to a measured difference of potential by means of a battery of small cells.[1]

§ 24. Measurement of the Potential Gradient

The potential gradient dV/dh may be found by measuring the difference of potential between two

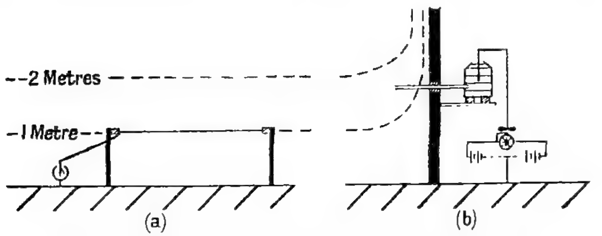

Fig. 11.—Measurement of the potential gradient.
(*a*) Absolute determination in open field.
(*b*) Continuous recording with the Kelvin water-dropper.

insulated conductors at different heights above the ground. Various forms of electrometer are employed, and one conductor may be the earth itself, while the other is a stretched horizontal wire about a metre above ground-level. Sometimes two wires at different heights are used. It is of course essential that the supports of the wires and the presence of the observer and his instruments should not materially alter the field to be measured. In the arrangement in Fig. 11 (*a*), the wire

is hung between amber or sulphur insulators at the ends of two vertical rods; its length should exceed twice its height above ground.

The wire when first set up is at a different potential to that of the air in its neighbourhood, and a current will flow between them which tends to remove this potential difference. The ordinary conductivity of the air is, however, too small for this process of equalisation to take place very rapidly and the electrometer reading will always lag behind a fluctuating field. It is necessary for this reason to place at the centre of the wire some active ionising agent which speeds up the process by increasing the local conductivity of the air. This may take the form of a glowing fuse of filter-paper, impregnated with a 5 per cent. solution of lead nitrate, or of a disc or spiral of metal coated with a deposit of ionium or radio-thorium. Such "collectors," as they are called, have the advantage of causing the wire and the electrometer to acquire the potential of the air at the centre of the stretched length, where the electric field is least disturbed by supports and observer. The time taken for equalisation of potential depends upon the type and activity of the collector and varies from 1 to 60 seconds. There is always some uncertainty as to the exact position at which the ionising agent exerts its effect, especially if a wind is blowing, but this can be avoided by making observations at different heights.

§ 25. Continuous Recording of the Potential Gradient

An adaptation of the arrangement just described is employed to obtain continuous records of the potential gradient, the electrometer being housed in a building and the collector carried on an insulated rod projecting from the wall. The earth's field is very much distorted by the presence of the building, and a reduction factor for the installation must occasionally be determined by making simultaneous measurements in an open field. Fig. 11 (*b*)

shows the nature of the distortion of the equipotential surfaces due to a building.

The Kelvin water-dropper is still largely used at recording stations as a potential equaliser. A jet of water from an insulated cistern within the building escapes from the end of a pipe passing through and insulated from the wall, and breaks up into fine drops at the point where the potential is required (Fig. 11). On starting the jet, the potential of the insulated system will in general be somewhat lower than that of the air at the end of the pipe; this end will carry an induced negative charge and the cistern a positive one. Each drop carries a negative charge away and leaves the system slightly higher in potential than before. Ultimately the negative charge at the end of the pipe disappears and the system is at the potential of the air at the point at which the drops break away. In practice this takes a time of the order of 30 seconds.

Radioactive collectors are also employed, but whatever the type of collector, the insulated system is usually joined to the needle of a quadrant electrometer of low sensitivity, and opposite pairs of quadrants connected to a battery whose centre is earthed. The record can be made intermittently, by means of a clockwork arrangement which depresses an inked pointer attached to the needle (Benndorf), or continuously, by receiving the spot of light from a small mirror on a moving strip of bromide paper.

§ 26. Variation of the Potential Gradient with Height above the Ground: Space-charge

A good many observations of the potential gradient have been made with balloons fitted with collectors of the radioactive or the glowing fuse pattern; they all indicate a rapid decrease in the field with increasing height above the ground (Table V, § 33). Though no great accuracy can be claimed for these measurements —for they are subject to errors arising from the dis-

tortion of the field by free and induced charges on the balloon and are but momentary observations of a quantity subject to considerable variation—it is well established that at a height of 10 kilometres the field has fallen to less than 1/50 of its value at ground level and is diminishing still further. Some recent experiments of Idrac,[2] in which a thermionic valve was employed as a voltmeter, suggest that at very great heights the gradient may sometimes be reversed in direction, but the significance of these results is not yet clear.

This rapid diminution in the potential gradient with height indicates that a free positive charge, practically equal to the negative charge on the surface of the earth, resides within the lower 10 or 15 kilometres of the atmosphere. Integration of the values of dV/dh obtained in balloon ascents gives a total potential difference between the earth and the air at a height of 15 kilometres of about 3×10^5 volts; at this height, and above it, the conductivity of the air is so great that this figure may be taken to represent the potential of the highly conducting "Kennelly-Heaviside layer" some 80 kilometres above the ground.

The density, ρ, of the positive "space-charge" referred to can be determined from Poisson's equation, which, if the lines of force are vertical, takes the form $\frac{d}{dh}(dV/dh) = -4\pi\rho$. Several investigators have endeavoured to examine the nature of the space-charge quite close to the ground, using collectors at different heights. The results obtained are conflicting as to the magnitude, and even the sign, of ρ in this region; evidently local conditions are of main importance.

§ 27. Variations of the Potential Gradient with Locality, Time and Season

Before passing on to the more important variations of the potential gradient, mention should be made of the fact that incessant fluctuations are observed over

short periods of time; these are due to local changes in the space-charge and the conductivity of the air in the vicinity of the collector.

For many years continuous records of the potential gradient have been taken at certain stations in the Northern Hemisphere. Less extensive information is available for the Southern Hemisphere, but observations have been made over considerable periods in the Arctic and Antarctic regions. Some very important measurements over the oceans have been carried out by the survey ship *Carnegie* of the Carnegie Institute of Washington. It appears that the sign of the potential gradient in fine weather is positive all over the earth's surface, but that over land areas its value varies considerably with local conditions. The mean of dV/dh at Kew is 317 volts/metre, while that at Davos in Switzerland is but 64 v./m. Over the oceans a value of 126 v./m., which varies little if at all with geographical position, has been found. The average value for the whole earth is not far from 120 v./m.

It would seem that the potential gradient observed on land is not a quantity of much fundamental significance, its value depending primarily upon local conditions of atmospheric conductivity. This is evident from the fact that a simple periodic variation according to *local* time is the basis of all fine-weather land records. The minimum values are obtained in the early morning at about 4 a.m. and the maxima in the evening between 6 and 8 p m. In many places a second maximum at 8 a.m. and a second minimun at midday are found. The amplitude of these daily variations at land stations sometimes reaches 50 per cent. of the mean value for the day.

The local time variation has been shown by Whipple [3] to be closely correlated with the amount and distribution of the smoke pollution of the atmosphere near large cities, where these continuous records have generally been made. Fig. 12 shows the variation at Kew of the potential gradient during the summer months for

the two periods 1898-1915 and 1916-28, together with the variation in the amount of pollution (determined by aspirating the air through filter-paper) for the period 1921-28. The figure shows not only the close relation which exists between potential gradient and pollution but also the striking effect of the introduction of "summer time" into Great Britain in the year 1916, after which the morning minimum and maximum moved back by approximately one real hour. The atmospheric pollution curve has maxima and minima which have been ex-

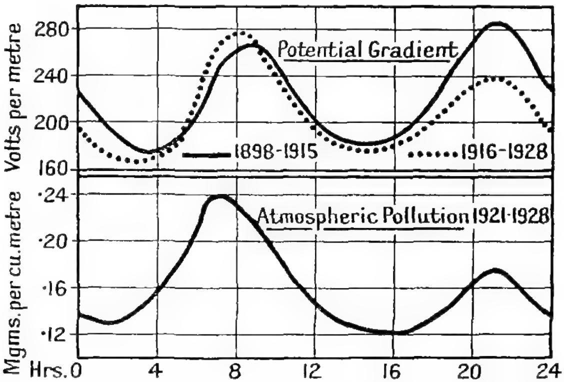

FIG. 12.—Relation between potential gradient and atmospheric pollution (Whipple).

plained by Simpson as due to the combined effect of variations in the amount of smoke produced in the city and variations in the stability of the atmosphere, that is in the mixing action of surface winds and general turbulence of the lower air.

As regards the annual variation of the potential gradient, land stations in both hemispheres show a maximum in the local winter and a minimum in the local summer. The only significant exception is the Antarctic region, where a reversal of phase occurs, giving rise to a maximum in the local summer and a minimum in the winter.

In view of the local origin of the variations of the gradient found at land stations near towns, observations over the oceans, where no pronounced local effects are to be expected, are of very great importance. Mauchly [4] has analysed the measurements of the *Carnegie* cruises and found that there is a well-marked diurnal variation at sea, with maxima and minima occurring at the same moment in all parts of the globe. Thus the potential gradient in all oceans is found to be about $15\frac{1}{2}$ per cent. below the mean at 5 hours, Greenwich Mean Time, and

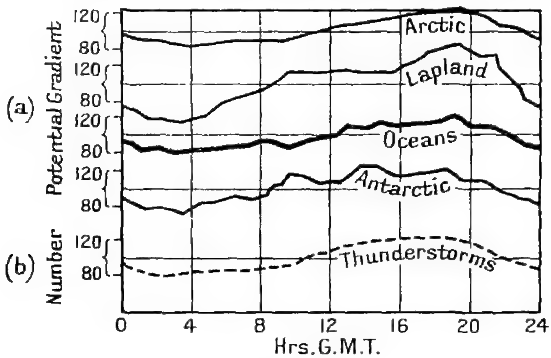

FIG. 13.

(*a*) Variation of potential gradient with universal time.
(*b*) Variation of thunderstorm activity over the earth's surface (Whipple).

about 20 per cent. above it at 19 hours, G.M.T. It is remarkable that the same 24-hour wave, progressing according to universal solar time, has been observed by expeditions to the Arctic and Antarctic regions and to Lapland. Fig. 13 (*a*) (after Whipple) illustrates these results, which have been confirmed by the last cruise of the *Carnegie*. In each case shown in the figure, the mean potential gradient has been reduced to 100 v./m. No appreciable annual variation has so far been observed over the sea.

Mauchly's discovery of a variation of potential gradient according to universal time is generally accepted as being of fundamental significance. The variation must clearly be ascribed to an alteration in the difference of potential between the upper conducting layers of the atmosphere and the earth; it affords an important means of testing any theory of the origin and maintenance of this potential difference.

§ 28. The Fine-weather Current: Conduction and Convection Currents

In the ordinary fine-weather field positive ions are driven towards the earth and negative ions away from it; the motion of these ions constitutes a downwardly directed conduction current. We have seen (§ 3) that the conduction current passing through 1 sq. cm. of a horizontal plane is given by $I_0 = F(\lambda_+ + \lambda_-)$, where F is the strength of the electric field and λ_+ and λ_- are the polar conductivities of the air at the point in question.

There is, however, a fine-weather current of a different nature which plays a part in the transfer of electricity by the atmosphere. If the air should contain at any point an excess of ions of one sign (a space-charge), movement due to wind or ordinary turbulence will give rise to a mechanical transference of electric charge. Thus if v is the upward vertical component of the velocity of the air and ρ the space-charge per c.c., the upward convection current due to this cause will be $v\rho$ per square centimetre. The real current-density in fine weather is thus the resultant of the downward conduction and the upward convection currents, and is given by $I = I_0 - v\rho$.

The mean value of the total fine-weather current is not far from 2×10^{-16} amperes per square centimetre or 2 microamperes per square kilometre, so that the total current flowing in this way between the upper atmosphere and the whole earth is about 1000 amperes. It would seem, from measurements which have been made in

various parts of the world, that this current varies much less with changes in geographical position, in time of day and season of the year, than does the potential gradient. Although in the past it has not received a great deal of attention, it would appear to be a more fundamental quantity than the potential gradient and less affected by purely local conditions.

§ 29. Direct Determination of the Conduction Current at the Surface of the Earth

Direct measurements of the magnitude of the conduction current which flows into the earth's surface have been made with the Wilson universal electrometer, with the test-plate mounted flush with the surface of the ground, as described in § 23. The curve of Fig. 14

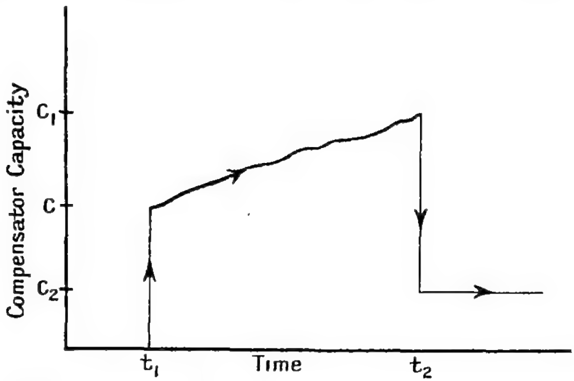

Fig. 14.—Measurement of conduction current with the Wilson universal electrometer.

illustrates the procedure followed in the measurements. At time t_1 the test-plate is exposed to the earth's field by removing the cover, and the capacity of the compensating condenser is adjusted to the value C necessary to bring the gold leaf back to the zero position. The

charge on the plate is then $-$ CV and the field strength $F = 4\pi\frac{CV}{A}$, where, as before, A is the area of the plate and $-$ V the potential of the outside of the compensator. The test-plate is now left exposed until time t_2 and kept continuously at earth potential by increasing the compensator capacity to balance the charge received from the positive ions driven from the air on to the plate. Owing to variations in the strength of the field, and so in the induced charge on the plate, this increase is not a regular one. At time t_2 the plate is again shielded with the cover and the compensator capacity is reduced from C_1 to C_2 to maintain the shielded system at earth potential. At this moment the field strength is $4\pi(C_1 - C_2)V/A$. Since at the beginning of the measurement the compensator capacity was zero, it is evident that during the interval, $t_2 - t_1$, the plate received a positive charge C_2V and the average value of the conduction current per square centimetre was $C_2V/A(t_2 - t_1)$. For routine observations the instrument is mounted on a tripod above the ground, and the readings are reduced to standard conditions by occasional determinations of the reduction factor, from observations in a pit.

§ 30. Indirect Determination of the Conduction Current at a Point in the Air

The conduction current at a point in the air can also be determined by separate measurements of the field strength, F, and the polar conductivities, λ_+ and λ_-. The methods used to determine F, or the potential gradient, have already been described in this chapter; the conductivities are usually found by the Gerdien method sketched in § 2. In another form of the Gerdien method, due to Schering, the cylindrical condenser is replaced by a large earthed wire cage, along the axis of which is a charged insulated wire connected to an electrometer. The rate at which the charge is dissipated by ions of opposite sign is determined in exactly

the same manner as in the Gerdien method, and the same equations apply if certain conditions are fulfilled. The cage must be large enough, and the charge on the wire small enough, to ensure that no considerable alteration in the number of ions is caused by the flow of current. The arrangement must be set up in the open air to ensure that ordinary atmospheric circulation prevents the development of the electrode space-charges to be described in the next section. And finally the whole cage must be shielded from the earth's field by a rough roof or by trees, otherwise it will carry an induced negative charge on the outside and rob the air entering it of negative ions.

§ 31. The Comparison of the Direct and Indirect Methods: Electrode Space-charge

The direct method due to Wilson measures the conduction current entering the earth; the indirect method measures the current at a certain height above the earth. In the latter case the current is carried by two oppositely moving streams of ions of opposite sign; in the former it consists of a single stream, for while positive ions enter the earth, negative ions do not pass from it into the atmosphere. The two methods, therefore, do not measure the same thing. One determines the true conduction current at a height, and the other its positive component at ground-level.

Consider a vertical cylinder of perfectly still air whose height is h and cross-section 1 sq. cm., with its base on the ground (Fig. 15). Let F′ and F be the electric field strengths at top and bottom of this cylinder, and λ'_+, λ'_- and λ_+, λ_- the polar conductivities. The conduction current through the top is $F'(\lambda'_+ + \lambda'_-)$; a positive charge $F'\lambda'_+$ enters, and a negative charge $F'\lambda'_-$ leaves, in each second. Through the bottom there is only a downward flow of positive ions, carrying away a positive charge $F\lambda_+$ per second. The cylinder as a whole thus

loses a positive charge $F\lambda_{+} - F'\lambda'_{+}$ per second, and a negative charge $F'\lambda'_{-}$ per second. This may be expressed as a development of a positive space-charge at a rate $F'\lambda'_{-} - (F\lambda_{+} - F'\lambda'_{+}) = F'(\lambda'_{+} + \lambda'_{-}) - F\lambda_{+}$ per second. The growth of this "electrode" space-charge will automatically decrease the field F' and will cease only when F' has such a value that as many negative as positive ions leave the cylinder in each second, i.e. when

$$F'(\lambda'_{+} + \lambda'_{-}) = F\lambda_{+} \quad . \quad (10)$$

The two sides of this last equation are respectively the conduction currents at height h and at ground level. Thus, for perfectly still air, the two methods we have described should give the same numerical values, though they measure different quantities. It appears probable that this conclusion is not realised in practice and that the Wilson method yields much smaller values than the other. Some average figures from the more extensive series of measurements are shown in Table IV.

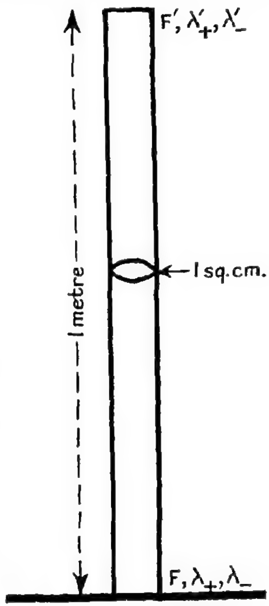

FIG. 15.—Electrode space-charge.

Still another consequence of the argument we have considered fails to appear in practice. Watson [1] has shown that during fine weather at Kew there is no sign of any electrode space-charge within the first metre above the ground, and that F' and F are the same.

As there can be no doubt that a space-charge would develop if the air were perfectly still, it has been

TABLE IV

MEASUREMENTS OF THE AIR-EARTH CONDUCTION CURRENT

Indirect Method.		Direct Method.	
Station.	Current (amps./sq. cm.).	Station.	Current (amps./sq. cm.).
Potsdam .	$2{\cdot}2\ 10^{-16}$	Kew . .	$1{\cdot}1\ 10^{-16}$
Davos .	2·0 ,,	Munich (Lutz)	1·0 ,,
Gottingen .	2·7 ,,		

suggested that ordinary atmospheric turbulence, by continually mixing the lower layers of the air, carries the excess positive charge upwards as rapidly as it is formed. If $F' = F$ and $\lambda'_+ = \lambda_+$, the positive charge developed in the cylinder per second is $F'\lambda'_-$. This must be removed by the air at such a rate that an upward connection current $v\rho = F'\lambda'_-$ per sq. cm. crosses the top of the cylinder, and the total resultant air-earth current at the height h is

$$I = I_0 - v\rho = F'(\lambda'_+ + \lambda'_-) - F'\lambda'_- = F'\lambda'_+ \simeq F\lambda_+.$$

If this view is correct, the quantity $F\lambda_+$, which is the portion of the conduction current measured by the Wilson method, is actually equal to the total current, conductive and convective, passing between the atmosphere and the earth. While this result may not hold in practice with great exactitude, the arguments outlined above indicate that the test-plate method gives a closer approximation to the total fine-weather current than the indirect method which determines the conduction current only. The test-plate result is the more valuable because it appears that any direct estimate of the convection current by itself is very difficult.

§ 32. Variations in the Fine-weather Current

Both direct and indirect methods of studying the air-earth current indicate that in fairly fine weather it is less subject to variations than is the potential gradient. Simultaneous measurements of the field and of the total conductivity of the air, $\lambda = \lambda_+ + \lambda_-$, show that these quantities vary to some extent in inverse ratio, with the result that their product, the conduction current, is less variable than either. Factors such as an increase in the number of condensation nuclei or the presence of fog particles, which decrease λ by raising the proportion of the slow ions present, cause a corresponding increase in the potential gradient. The simple form of the daily variation in the earth's field has its counterpart in an inverse variation of the conductivity, with a maximum in the early morning between 3 and 4 a.m. The extreme effect of atmospheric pollution near large cities, which increases the potential gradient and diminishes the conductivity, has already been discussed (§ 27). In general the reciprocal relation between λ and F is not exact enough to keep the conduction current invariable. At land stations in Europe it is found to have an annual variation which coincides in phase with that of the potential gradient but has a smaller amplitude. At most stations its daily variation follows the changes in the conductivity.

As shown by Table IV the mean values obtained at European stations for the conduction current, using the indirect method, all lie close to $2{\cdot}0 \times 10^{-16}$ amps./sq. cm. The observations over the sea and near the poles give about double this value. Since, as we have seen, to make allowance for the upward convection current near the ground these figures must probably be reduced by half, the total air-earth current in fine weather may be taken to have an average value of 2×10^{-16} amps./sq. cm. over the whole globe.

§ 33. The Conduction Current at Great Heights

It has been found from simultaneous determinations of potential gradient and atmospheric conductivity during balloon ascents that the rapid decrease in F with height above the ground is accompanied by an increase in λ, of such magnitude as to keep the conduction current approximately constant. Table V shows the results of an ascent by Wigand,[5] in which it will be seen that when the field had fallen to 1/14 of its value at the ground, the conductivity had risen eleven-fold.

TABLE V

Conduction Current during Balloon Ascent (Wigand)

Height (metres).	F (volts/metre).	λ (e.s u).	I (amps/sq. cm).
0	136	$1{\cdot}1\ 10^{-4}$	$1\ 7\ 10^{-16}$
2500	27	4·8 ,,	1·4 ,,
4400	18	8·2 ,,	1·6 ,,
6500	8·8	12 6 ,,	1·2 ,,

§ 34. The Significance of the Air-earth Current

The general constancy of the air-earth current with height above the ground and in different parts of the world suggests that, at any rate to a first approximation, it has its origin in a constant potential difference between the conducting layers of the upper atmosphere and the earth, and that its actual value depends upon the resistance of the air between the two regions. Now the conductivity of the upper portion of this air-path is entirely due to the penetrating radiation * and is unlikely to alter at all with time of day, locality or season (§§ 20, 21). But below a height of about 2 kilometres over land

* This refers to the regions from 15 to 30 kilometres above the ground. The Heaviside layer 80 kilometres up has a solar origin.

areas, an additional ionising influence is exercised by radioactive matter in the air and the earth, which, as we have seen (§ 5), varies very much. Moreover, the conductivity of the air in this region is dependent upon the relative concentrations of the large and the small ions, and these alter with the number of nuclei present. The current-bearing column of air above the point of observation thus consists of an upper constant resistance and a lower variable one in series with each other. The extent to which the current carried by the air-column alters with local conditions will depend on the ratio which the lower variable resistance bears to the total resistance.

The conductivity above 14 kilometres is so high that we may, for the sake of illustration, suppose that this is the actual height of the upper positively charged layer and take the mean specific resistance of the air between it and the earth to be $7{\cdot}1 \times 10^{14}$ ohms-cm., the value found by Wigand at a height of 6·5 kilometres. Then the total resistance of an air-column of cross-section 1 sq. cm. would be 10^{21} ohms. If the variable part of the column be the first 100 metres above the earth, where the specific resistance is usually 8×10^{15} ohm-cm., this portion will contribute 8×10^{19} ohms to the total. We see that even if the conductivity of the first 100 metres were halved, the effect upon the air-earth current would amount to but 4 per cent. The potential gradient near the ground, on the other hand, would be doubled, for from this point of view it represents the fall of potential per unit length of the lowest part of the current-bearing column. It is evident that potential gradient measurements over land, and in particular near cities, cannot be expected to yield information which bears directly upon the question of the total potential difference between upper atmosphere and earth. In the case of observations at sea and in polar regions the position is different; practically all the ionisation throughout the whole column is then due to the penetrating radiation, and the only local effect which can influence the

fine-weather gradient is a change in the number of nuclei available for the capture of small ions.

§ 35. The Electric Field during Disturbed Weather

The normal downwardly directed field, due to a positively charged upper atmosphere and a negatively charged earth, is frequently disturbed when the weather ceases to be fine. During fog, as explained in § 32, it is very much increased, and may reach ten times its normal value ; during dust-storms, of the type common in semi-arid regions and deserts, powerful reversed or negative fields are usual and reach 10,000 volts per metre. The effects found during cloudy weather and rain are variable, and range from gradients of the order of a few hundred volts per metre in a fine drizzle to as much as 50,000 volts per metre under thunderclouds.

Generally speaking, negative fields predominate when light or steady rain is falling, though occasional positive excursions are observed. In the case of heavy rainstorms and thunderclouds, the sign of the field depends upon the portion of the cloud passing over the point of observation, but here, too, there is evidence that negative gradients are the more frequent.

The field during thunderstorms fluctuates a great deal if the cloud is active in producing lightning. Its value at the ground does not as a rule exceed 10,000 to 20,000 volts per metre except for very brief intervals occupying fractions of a second. Equally high fields are observed during heavy rainstorms unaccompanied by lightning. During a snowstorm the field is usually positive and in a heavy storm may attain 10,000 v./m. The methods employed for the study of the large and rapidly varying fields of thunderclouds are described in Chapter IV.

§ 36. The Charge on Rain and Snow ; the Precipitation Current

The apparatus required to examine the charge carried to the earth by precipitation—rain, hail, snow, and

sleet—is comparatively simple, and consists essentially of an insulated metal vessel to catch the rain, etc., and an electrometer to measure the charge received. Great precautions must, however, be taken to eliminate the disturbing effect of the negative charge given to the air as the drops splash on the container (Lenard effect, § 43). If some of this is lost a spurious positive charge will be indicated. For a similar reason, the entry of spray due to the splashing of rain falling outside the receiving vessel must be prevented. The receiver must be screened electrostatically from the electric field outside. The rate of rainfall, and sometimes the average size of the drops, is also observed; with sensitive forms of the apparatus the charges carried by individual drops may be measured. As the charges brought down vary considerably, it is necessary in all such work to carry out a long series of observations.

Since the year 1908, when the modern series of observations began, all observers agree in finding a preponderance of positive charge on rain in general. The average ratio of the quantities of positive and negative electricity brought down by all kinds of precipitation varies according to the situation of the observer. Some of the values found for this ratio are: Potsdam (Schindelhauer), 1·4 : 1; Puy-en-Velay (Baldit), 2·4 : 1; Dublin (McClelland and Nolan), 4·8 : 1; Simla (Simpson), 2·4 : 1; Otago, New Zealand (Marwick), 3·2 : 1. As exceptions to the general rule, it may be noted that snow usually carries a considerable negative charge, and that the fine drops of a drizzle are also negative. Simpson showed in 1908 that thunderstorm rain—the most heavily charged of all forms—is positively charged when it falls from the front of the cloud and negatively charged elsewhere. Ordinary steady rain, unaccompanied by sudden changes in the barometer, shows the most regular as well as the highest excess of positive charge. If such rainfall alone is considered, Baldit's average ratio of 2·4 : 1 for all types of precipitation rises to 4·3 : 1.

The charge per c.c. in the case of heavy thunder-

storm rains reaches average values of 30 to 40 e.s.u., while single drops may carry 100 to 200 e.s.u. (Gschwend) and attain potentials as great as 300 volts. The mean values reported by different observers for the nett charge per c.c. on all forms of precipitation during a long series of observations range from + ·029 e.s.u. at Potsdam (Schindelhauer) to + ·176 e.s.u. at Simla (Simpson).

The transfer of electricity from atmosphere to earth in this manner constitutes a "precipitation" current which appears on the whole to be in the same direction as the fine-weather conduction current, since it carries more positive than negative charge to the earth. In the case of thunderstorms this current sometimes reaches values as high as 2×10^{-11} amps./sq. cm., and the total convection current carried by the rain in such a storm may amount to more than 0·1 amp. Before it is possible to frame a trustworthy estimate of the amount of the excess of positive over negative electricity conveyed to the whole earth in this way, much more information is required from tropical and equatorial regions.

§ 37. Currents due to Point-discharge

It is well known that an electric glow or brush discharge takes place from a pointed conductor when placed in a sufficiently strong field. It consists of a stream of ions of the same sign as the charge upon the point. Discharges of this kind must, therefore, be expected from the ends of exposed conductors on the surface of the earth whenever the potential gradient and the height and sharpness of the ends are sufficiently great. Indeed the earliest investigators, Franklin with his kite, Dalibard and Lemonnier with their high metal rods, detected electric fields, the first two in thundery, and the last in fine weather, by means of this current. Under favourable conditions, afforded by mountain peaks or the masts of ships exposed to the intense

fields of a thundercloud, the glow discharge may become very conspicuous (St. Elmo's fire).

It was first pointed out by C. T. R. Wilson [6] that point-discharge currents of much smaller intensity must play an important part in the interchange of electricity between the earth and the atmosphere. The fields prevailing under rainstorms and thunderclouds very frequently reach values sufficiently great for considerable currents to be discharged from exposed conductors, such as trees, bushes, housetops and even fields of grass. A conductor need not end in a sharp point or project to a great height in order that it should begin to act as a discharger. For example, an earth-connected sphere of radius 1 cm. need only be raised to a height of 3 metres in the not unusual field of 10,000 volts per metre for the electric intensity at its surface to reach the critical value of 30,000 volts per cm. at which brush discharge begins. A practical instance of the correctness of this suggestion is afforded by some experiments made by the author, in which a small tree, 4 metres high, was cut off at its base, mounted on insulators and connected to earth through a galvanometer. Exposure to the fields due to nearby thunderstorms yielded the average point-discharge currents shown below:—[7]

TABLE VI

POINT-DISCHARGE CURRENTS FROM A SMALL TREE (SCHONLAND)

Field (Volts/metre).	Current (Microamperes).
− 3,500	0·07
− 5,500	0·20
− 11,000	1·00
− 16,000	4·00

The contribution of a single tree such as this must be multiplied by a very large factor to obtain the total current flowing between the earth and a thundercloud; in the case quoted, the tree was typical of the exposed natural conductors for many miles around, and a rough surface integration over the area affected by the cloud indicated that the total point-discharge effect was of the order of 2 amperes in an upward direction.

Since the direction and magnitude of the large fields associated with disturbed weather vary considerably, it is extremely important to know the relative amounts of positive and negative electricity discharged from an exposed point over a long period of time. In a series

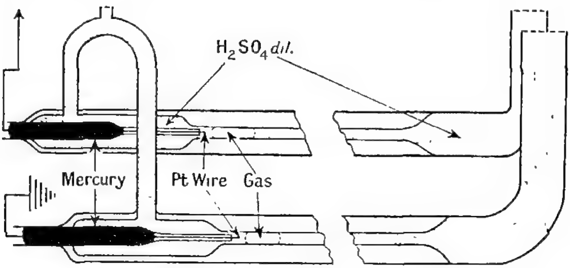

FIG. 16.—Water micro-voltameter (Wormell).

of observations by Wormell,[8] a sharp point carried on insulators at the top of a pole was connected by cable to one electrode of a voltameter filled with dilute sulphuric acid, the other electrode being earthed. The voltameter (Fig. 16) was constructed of fine capillary tubing (diameter 0·8 mm.) and the gases evolved were separately collected, their volumes being determined from the lengths of the bubbles formed. If v_1 and v_2 are the volumes of the mixed gases collected at the earthed and the point-connected electrodes respectively, the quantities of positive and negative electricity, q_1 and q_2, which have been discharged from the point can be separately determined in the following way: If one unit

of quantity liberates in electrolysis c c.c. of oxygen and $2c$ c.c. of hydrogen, we have

$$q_1c + 2q_2c = v_1 \quad \text{and} \quad 2q_1c + q_2c = v_2,$$

from which we may determine q_1 and q_2.

The quantities discharged in this way have been examined by Wormell at Cambridge over a period of three years. Some of his results are shown in the table below :—

TABLE VII

INTEGRATED EFFECTS OF POINT-DISCHARGE (WORMELL)

Year	1927	1928.	1929.
Positive quantity discharged, q_1 (coulombs).	·28	·24	—
Negative quantity discharged, q_2 (coulombs).	·14	·11	—
Nett positive discharge, $q_1 - q_2$ (coulombs) .	·14	·13	·11

Each year shows a considerable nett loss of positive electricity by the discharger.

In a more detailed discussion it is noted that of 147 separate periods of disturbed weather accompanied by precipitation, 103 showed an excess of upward current (+ ve discharge), 34 an excess of downward current (− ve discharge), and in 10 the nett discharge was zero. This is what would be expected from the predominance of negative potential gradients in disturbed weather. It is to be hoped that similar measurements will be made in other parts of the world.

§ 38. The Interchange of Electricity Between the Earth and the Atmosphere; the Maintenance of the Earth's Charge

The negatively charged earth and the positively charged upper layers of the atmosphere form two plates of a spherical condenser with the lower air as the dielectric. Although the conductivity of the air between the plates is small, the applied potential difference is great enough to make the leakage of charge through it very considerable. The average value of the charge per unit area of the earth's surface is 9×10^{-14} coulombs/sq. cm., and that of the fine-weather conduction current is 2×10^{-16} amps. per sq. cm. Left to itself, a condenser of this kind would be discharged by internal leakage in a time of the order of ten minutes. But this is not all: to the dissipation of the earth's charge by the fine-weather current we must add the effect of the charge conveyed by rain and snow, for it has been shown that on the whole the nett charge conveyed to the earth by precipitation is positive in sign. Thus to the conduction current of 1000 amperes over the whole surface of the earth must be added a precipitation current in the same direction, estimated by Wigand [9]—though any such estimate can only be provisional—at 400 amperes, making a total of 1400 amperes tending to dissipate the 500,000 coulombs with which the earth is charged.

In spite of the continuous operation of these two factors, the earth's charge remains practically constant, so it is clear that there must exist some reverse or compensatory process which neutralises their effect. Some agency must be continuously at work replenishing the charges on the earth and in the upper air. The discovery of the mechanism responsible for this replenishment has been one of the chief problems of atmospheric electricity, and many suggestions have been made to this end. Two of these, though they have not proved acceptable, will be briefly mentioned.

The suggestion of Ebert is based upon the experimentally verified view that ionised air escaping through the narrow capillaries of the earth's crust is richer in positive than in negative ions, for the higher mobility of the negative ion causes it to diffuse more rapidly to the walls of a narrow passage. The air escaping from the earth and ionised by radioactive emanations, etc., thus gives a positive charge to the lower atmosphere, which Ebert suggested was conveyed to considerable altitudes by convection currents. It appears, however, that the Ebert effect in the lower atmosphere is actually extremely small and that the upward air-currents are quite insufficient to produce the necessary convection.

Another suggestion, made by Simpson and developed by Swann, is that negative charge is conveyed to the earth by fast-moving negatively charged corpuscles or β particles. This view meets with the difficulty that the ionisation produced in the air by the number of β particles needed to maintain the earth's charge would be very large, many times greater than that observed and known to be due to other causes. The objection can be avoided only by ascribing a very small ionising power to an extremely fast β ray (Swann). This is not in accord with what is known of the fastest β rays from radioactive substances, whose ionising powers vary very slowly with their energies.

It was first suggested by C. T. R. Wilson that it is to regions of disturbed weather that we must look for the mechanism of replenishment, that under rain and thunderclouds, where the potential gradient is more often reversed than in the normal direction, there are two processes at work which convey considerable quantities of negative charge to the earth. The first of these is the action of point-discharge from conductors projecting from the ground. This has already been discussed, and it has been shown that experimental tests indicate that it is of great importance. The second process is the charge conveyed to the ground by lightning flashes from thunderclouds. Every second, as will be

seen in Chapter IV, some 100 lightning discharges occur over the whole surface of the earth, each involving the passage of a charge of the order of 20 coulombs. Only a fraction of these, perhaps one in four on the average, strike the ground. The effectiveness of this process of transference of charge will depend upon whether more charges of one sign than another are conveyed to the earth. If, for example, practically all the charges transferred were negative, the process would be equivalent to a continuous compensation current of about 500 amperes. Though there is evidence that this is actually the case, the question is still under investigation.

An estimate of the annual electrical "balance-sheet" of a square kilometre of ground at Cambridge has been made by Wormell,[8] on the basis of his point-discharge measurements and observations on lightning discharges. The air-earth conduction current loss and the effect of rain and snow were estimated from average values of these quantities. The results were :—

	Coulombs/sq. km /annum.
By natural point-discharge, gained .	−100
By lightning, gained	− 20
By atmospheric conduction, lost . .	− 60
By precipitation, lost	− 20
Nett gain of negative charge . . .	− 40

Though such an estimate is only a very rough one, it appears to be quite possible that in this locality the four processes approximately balance one another, or even that the earth gains a negative charge.

This suggestion of Wilson involves more than the interchange of electricity between the bases of cumulo-nimbus clouds, showers and thunderstorms, and the ground; he regards them as equally active in supplying positive electricity to the upper air. On this view the cloud acts as an electrical generator which removes positive electricity from the earth and supplies it to the conducting layers of the atmosphere above, by which it is

rapidly distributed in such a way as to maintain them at a constant potential of about 3×10^5 volts, in spite of losses.

If this view is correct, we should expect the potential difference, and so the potential gradient, to show a maximum when the thundery regions of the earth are at their maximum activity.[10] The diurnal variation of the gradient discovered by Mauchly (§ 27) should agree in phase with the diurnal variation of thunderstorm activity over the earth. It has recently been shown by Whipple[11] that this appears to be the case. Whipple's examination of the diurnal variation of the world's thunderstorms is represented by the bottom curve of Fig. 13 (*b*), where it will be seen that the parallelism with the variation of the potential gradient over the oceans and the polar caps is indeed very close.

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CHAPTER IV

THE ELECTRIFICATION OF THUNDERSTORMS

§ 39. The Number of Thunderstorms over the Globe

WHILE a thunderstorm is a comparatively rare event at any one station in temperate latitudes, the number of thunderstorms per day experienced by the earth as a whole is very large. Even in such a limited area as France there are comparatively few days in the year where thunder is nowhere reported. The total amount of thunderstorm activity which takes place at various seasons of the year has been determined by Brookes [1] from information taken from all parts of the world. According to his estimate, the earth experiences 16,000,000 thunderstorms per annum, or 44,000 per day. Taking one hour as the average duration of each—a modest estimate—there will on the average be 1800 thunderstorms in progress in different parts of the world at any one moment. In terms of lightning flashes this would mean that something like 100 flashes occur over the earth in every second, or 360,000 per hour.

It appears from recent measurements (§ 42) that the average thunderstorm develops and expends two or three million kilowatts of electrical energy continuously during its hour or two of activity. Over the whole earth, therefore, thunderstorms are continuously expending electrical energy at a rate of some 4×10^9 kilowatts. This figure is so large that by the employment of even a small fraction of its power in an apparently

insignificant "side-line," the thunderstorm could be of importance in a number of geophysical problems. Amongst these may be mentioned the generation of Hertzian waves (atmospherics), the maintenance of the earth's negative charge and of the positive charge on the upper air (§§ 37 and 38), the production of high-speed electrons (§ 21), and of ionisation in the upper atmosphere.

§ 40. The Measurement of the Electric Fields and Field-changes due to Thunderstorms

Practically all the quantitative information on the electrification of thunderstorms which has so far been obtained has been derived from measurements of the electric fields and field-changes produced by them at the surface of the earth. The development of these measurements and the interpretation of the results obtained is mainly due to Professor C. T. R. Wilson.[2]

The general principle is similar to that of his test-plate or induced charge method of studying the fine-weather field (§ 23), in which an exposed conductor is kept at zero potential by alteration in the capacity of a compensating condenser, when the charge induced upon the conductor by the field is equal to that given by the compensator to the system. In the measurement of rapidly changing thunderstorm fields, however, compensation is produced automatically and practically instantaneously by means of an ingenious form of capillary electrometer. As shown in Fig 17, this consists of a small bubble of dilute sulphuric acid enclosed between mercury threads in a narrow capillary tube. The threads lead to end-cups of mercury, and one of these, A, is connected to a test-plate T insulated from, but flush with, the surrounding ground, while the other is connected to the earth. Between the mercury and the glass there is everywhere a thin film of dilute acid, so that an electrical "double layer" surrounds the mercury and makes each half of the electrometer a condenser charged to the potential difference of the double

layer, about ·90 volt, the mercury being positively charged with respect to the acid. If a positive charge is given to the left-hand side, it will increase the charge on the left-hand condenser and the left-hand thread will move forward a certain distance so as to absorb this charge, by increasing the area of the acid-mercury surface and the capacity of the left-hand condenser. A corresponding movement of the right-hand thread causes a reduction in area on this side and the passage to earth of a quantity of electricity exactly equal to that originally supplied to the electrometer. The distance moved, y, is easily seen to be related to the charge passed through, q, by the equation

$$q = 2\pi r C V'(1 - r/R) y,$$

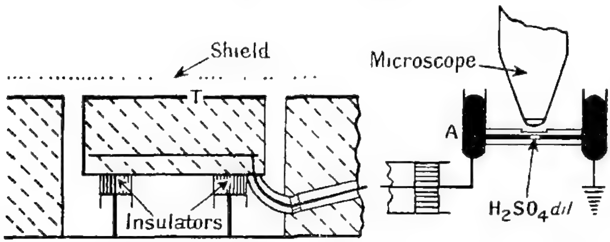

FIG. 17.—Measurement of the electric fields and field-changes due to thunderstorms (C. T. R. Wilson).

where r is the radius of the capillary and R that of the end-tube, V′ the potential difference, and C the capacity per unit area, of the double layer. Thus $q = ky$, where k is an instrumental constant which can be determined by discharging a known quantity of electricity through the instrument. The arrangement thus gives a linear relation between the movement of the acid bubble and the quantity discharged. In spite of the large value of C, about 30 microfarads per square centimetre, it is extremely sensitive. By using a microscope focussed upon one end of the bubble quantities of the order of 10^{-9} coulombs may be measured. A permanent record

is best obtained by placing a slit and a moving photographic plate in the focal plane of the objective.

When connected to a test-plate the electrometer automatically maintains it at earth potential and records by its movements the quantities of electricity passing to and from the earth as a result of changes in the external field and of the induced charges on the plate. The type of record obtained from this arrangement during a thunderstorm is illustrated by the thick line of Fig. 18. An earthed cover is swung over the test-

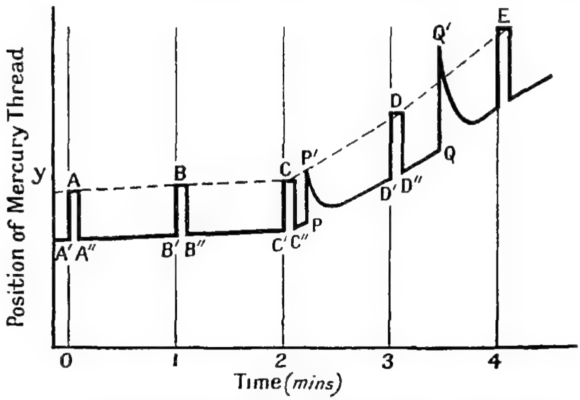

FIG. 18.—Capillary electrometer record during a thunderstorm.

plate for a few seconds at the beginning of each minute, causing the movements A′AA″, B′BB″, etc., as the induced charge flows away to earth and then returns. The points A, B, C, E would lie on a horizontal line if there were no actual transfer of charge between the plate and the atmosphere taking place, that is to say, no conduction current, point-discharge, or transport of charge by rain. The first two minutes in the figure show the slight slope due to the conduction current only, the last two show the effect of a fall of positively charged rain. The charge, for example, conveyed to the

plate by rain and conduction during the third minute is $k(y_D - y_C)$ and the mean value of the current per square centimetre $k(y_D - y_C)/60A$, where A is the area of the plate and k the constant of the electrometer. The dotted line joining the points A, B, C, D, E gives the reading of the instrument when the test-plate is shielded at any time during the record, and this line serves as the zero from which the field and field-changes are measured. The field at the end of the first minute, for instance, is given by $F = 4\pi k(y_B - y_{B'})/A$.

Two sudden interruptions in the record, such as are caused by lightning flashes, are shown at P and Q. At P the record suddenly touches the dotted line, indicating that for a moment the field was completely destroyed and the original induced charge ran to earth through the electrometer. At Q the movement extends beyond the dotted line, showing that the field was actually reversed in sign by the discharge and reached a reverse value equal to half its original strength. The shape of the record immediately after the passage of these lightning flashes, the recovery curve as it is called, is of considerable interest; it will be seen in these two examples that the thundercloud fields and charges recovered their original values after the lapse of something like twenty seconds. The rate of recovery was approximately exponential, being most rapid immediately after the flash and becoming very slow as the cloud approached its original electrified condition. We may find the magnitude of these sudden field-changes by the same method used for the field itself; if Δy be the movement PP′, or QQ′, the sudden field-change which caused it is given by: $\Delta F = 4\pi k \,.\, \Delta y/A$.

The test-plate arrangement, using a plate 50 cm. in diameter, is suitable for the measurement of fields of strength exceeding 1000 volts/metre, such as are caused by thunderstorms when they come within a distance of about 8 kilometres. For more distant storms Wilson employs a different form of exposed conductor, a copper ball 1 foot in diameter mounted on insulators at the end

of an iron pipe 5 metres in length (Fig. 19). When the arrangement is in use the pipe is held vertical; by turning it about a horizontal axis through its base, the pipe is lowered and the ball enters an earthed metal case which shields it from the field. Connection to the capillary electrometer is made by an insulated wire passing down the centre of the pipe.

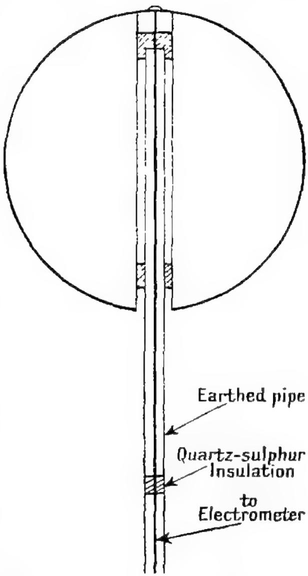

FIG. 19.—Elevated sphere for measuring fields of distant thunderstorms.

The field producing a given movement of the electrometer is calculated as follows: The charge q induced upon the ball when raised to a height h above the ground must be such as to maintain it at earth potential, for the ball is earth-connected through the electrometer. Thus if V is the undisturbed potential of the air at the point occupied by the centre of the sphere of radius r, we have $V + q/r - q/2h = 0$, the third term allowing for the effect of the electrical image of the charged sphere in the earth. If the movement of the electrometer caused by raising the ball from its case to the height h is y, we have $q = -ky$ and $V = k(1/r - 1/2h)y$. This observation therefore gives for the mean potential gradient between the ground and the height h,

$$F = V/h = (k/h)\,(1/r - 1/2h)y.$$

The same equation relates the sudden changes of field ΔF to the corresponding sudden movements Δy of the

electrometer. Since these relations, like those derived for the test-plate, are derived on the assumption of a perfectly plane earth, small corrections must be applied to allow for the effect of the concentration of the lines of force upon the hut used to house the recording instrument.

The effects obtained from the elevated sphere become too small for accurate measurement when the storm is more than 20 kilometres away; for still greater distances a wireless aerial has been used by some workers. Thunderstorm fields are of the same order as the fine-weather field, about 100 v./m., when the cloud is 20 kilometres off. At distances of 6 to 8 kilometres they often reach 5000 v./m., and when the storm passes overhead values of 10,000 to 20,000 v./m. are commonly observed, while the sudden changes may exceed these figures very considerably for a small fraction of a second. In an active thunderstorm the fields are constantly altering as a result of the neutralisation of part or all of the charge by lightning discharges, of the subsequent rebuilding of the destroyed charges, and of movements of the cloud as a whole.

§ 41. The Changes of Field Accompanying Lightning Discharges

The methods described have been used for the measurement of the field-changes due to lightning in England and in South Africa,[2, 3] and have led to valuable information as to the magnitude of the electrical quantities involved in the thunderstorm and the distribution of the charges in the cloud. The principles to be followed in the interpretation of the observations have been very completely discussed by Wilson.[2] Their application is not always an easy matter, for though the structure of most thunderclouds may be the same in principle, meteorological conditions lead to frequent departures from any simple model which can be framed. Anything like a full discussion of this question would occupy too

much space, and so we will limit ourselves to the detailed consideration of a type of lightning discharge which takes place within the thundercloud without reaching the ground and which is the most frequent kind of flash in a tropical storm. In the Cape Province of South Africa nine-tenths of the discharges are of this nature, and simply involve the downward movement of a positive charge from a height of 4 or 5 kilometres to one of 2 or 3 kilometres.

Consider a simple cloud-model in which a positive charge Q is distributed through a spherical region with its centre at a height H_2 above the ground. To calculate

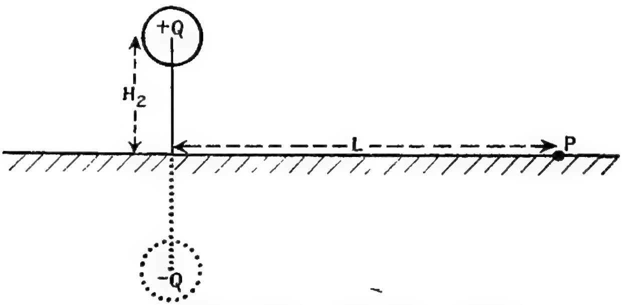

FIG. 20.—Thundercloud charge above the earth.

the resulting downwardly directed field at the earth's surface we may regard the charge induced on the conducting earth as replaced by the image-charge $-Q$ at a depth H_2 below the ground and we may disregard the curvature of the earth. The field at any point such as P (Fig. 20), at a horizontal distance L, is then that due to an electric doublet of moment $2QH_2$, and its strength is given by $F_2 = 2QH_2(H_2^2 + L^2)^{-3/2}$, vertically downwards. If a lightning discharge causes the positive cloud-charge to pass to a lower level H_1, the new field $F_1 = 2QH_1/(H_1^2 + L^2)^{-3/2}$ and the sudden field-change due to the discharge is

$$\Delta F = F_1 - F_2 = 2Q[H_1(H_1^2 + L^2)^{-3/2} - H_2(H_2^2 + L^2)^{-3/2}]. \quad (11)$$

It may be noted that this equation applies equally well if the act of lowering the charge Q involves the neutralisation of an equal and opposite charge $-Q$ at the height H_1, for the effect of adding the field due to this charge to both F_2 and F_1 disappears in computing the field-change ΔF.

Equation (11) shows that the change of field should be zero at a distance L_0 such that

$$H_1/(H_1^2 + L_0^2)^{3/2} = H_2/(H_2^2 + L_0^2)^{3/2},$$

that it is negative for $L > L_0$ and positive for $L < L_0$. This reversal distance L_0 is found to be 6·7 kilometres, if H_2 and H_1 are 6 and 3 kilometres respectively.

South African thunderstorms show this reversal in the sign of the majority of the field-changes very clearly. The following Table[3] gives the results of observations upon some fifty storms, giving 2800 observed field-changes which were grouped as "distant" when more than 15 kilometres from the station and as "near" when within 7 kilometres; in many cases the same storm was observed in both groups.

TABLE VIII

SUDDEN FIELD-CHANGES DUE TO THUNDERSTORMS (SCHONLAND)

Distant Storms.		Near Storms.	
Positive ΔF.	Negative ΔF.	Positive ΔF.	Negative ΔF
250	2375	188	9

The phenomenon of zero values for the field-changes, due to discharges at the reversal distance, can only be observed in isolated instances. In practice, owing to the discharge not taking a vertical but an inclined path, and for other reasons, the simple concept of a reversal point must be replaced by that of a reversal zone, in

which small field-changes of either sign are equally probable, and which, in the storms of Table VIII, may extend from 7 to 15 kilometres from the point of observation.

When the distance L is great compared with the heights H_1 and H_2 of the cloud-charges, equation (11) takes an important form ; it becomes

$$\Delta F = - 2Q(H_2 - H_1)/L^3 \text{ or } 2Q(H_2 - H_1) = - \Delta F \times L^3.$$

The quantity on the left is now the difference between the electric moments of the thundercloud charges before and after the discharge ; it is referred to as the electric moment of the discharge. All that is needed to determine it is a measurement of the field-change due to a distant lightning flash and of the distance L, which can be found from the interval between the flash and the thunder it causes.

Measurements of the electric moments of discharges have been made both in England and South Africa, with similar results. The average value found is about 2×10^{16} e.s.u. $\times$ centimetres, or 60 coulomb-kilometres, in Europe, and 90 coulomb-kilometres in the more elevated tropical storms. Direct observation in the latter case gives for the vertical lengths of cloud-flashes, $H_2 - H_1$, values lying between 1·5 and 3 kilometres, so that the average quantity of electiicity conveyed by a lightning discharge is about 20 coulombs. The usual range of variation for individual discharges is not large, from 2 to 100 coulombs if the length of the discharge is supposed constant, and probably much less if this quantity were actually observed in every case.

§ 42. Magnitudes of the Electrical Quantities Involved in Thunderstorms

The determination of the average quantity of electricity destroyed by a flash of lightning provides the starting-point for the evaluation of a number of important electrical quantities involved. A single flash, of the kind we have discussed, generally completely discharges

the whole cloud and momentarily reduces the field from it to zero. A fairly active cloud produces one flash every twenty seconds; to feed the flashes such a storm would have to generate electricity at an average rate of 1 coulomb per second, i.e. the mechanism would have to provide an average current of 1 ampere. Reference has already been made to the rapid rate at which the initial regeneration of the cloud-charges takes place. It is usually such as to be able to restore the whole electrification in about five seconds if it continued for that time. The exponential form of the recovery curves suggests that this initial rate may remain constant, but that, as the cloud charges grow, effects occur which tend to dissipate them and which are proportional to the quantity of charge present at any time. On this view the rate of regeneration involves a current of 20/5 coulombs/sec. or 4 amperes. The electrical energy generated by the storm is thus only partially employed in the feeding of lightning flashes; at the moment just before a discharge most of the power of the machine is being expended in overcoming various leakage effects. It may happen in this way that the field in the cloud never reaches the sparking value: this appears to be the case in many heavy rainstorms.

Before proceeding further it is necessary to know the critical value of the electric field strength at which the spark which we call lightning takes place. In air at ordinary pressure this is 30,000 volts per centimetre, but in a thundercloud the pressure is lower and the presence of small drops of water introduces a new and important factor. Macky[4] has recently investigated the behaviour of water-drops in strong electric fields, and finds that when exposed to increasing fields a drop of radius r cm. becomes elongated, until at a definite field strength, given in volts per cm. by $F\sqrt{r} = 3875$, it becomes unstable and filaments are drawn out from the ends. When instability occurs, a discharge passes through the drop in exactly the same manner as it would through a conductor pointed at both ends, with char-

acteristic luminous glows. The fields producing discharge through the drops are unaltered by reduction of the air-pressure over a wide range; for example, the field required to cause discharge through a drop 0·227 cm. in radius remains at 8400 volts/cm. as the pressure drops from 76 to 25 cm. of mercury. At still lower pressures the discharge is unaffected by the drop and the sparking potential falls in the usual way. The removal of water by the formation of filaments sets a limit to the size of the drops which can exist inside a thundercloud. Macky concludes from his experiments that no drops larger than 0·15 cm. in radius can be present in the fully charged cloud, and that the critical value of the field at which breakdown occurs is of the order of 10,000 volts/cm. or 33 e.s.u.

With this information the potential and the volume of the charged portion of the cloud can be estimated. Although the results might be expected to depend very much upon the manner in which the charge is distributed, it appears that whatever assumptions are made on this point the values obtained are of the same order of magnitude. Consider a simple spherical distribution of a charge q on a sphere of radius R. the electric field at the surface of the sphere is given by $F = q/R^2$, and at the moment at which a discharge takes place q is 20 coulombs or 6×10^{10} e.s.u. while F has the value 33 e s.u. Hence $R = 427$ metres and the dimensions of the charged region are of the order of 1 kilometre. The potential at the surface is given by $V = q/R = FR = 33 \times 4{\cdot}3 \times 10^4$ e s.u. or $4{\cdot}3 \times 10^8$ volts. If the radial electric force within the sphere everywhere reaches the sparking limit, the potential at the centre is 2FR or 9×10^8 volts. Two such spheres carrying opposite charges would therefore be at a difference of potential of the order of a thousand million volts.

A similar result follows if we make the very different assumption that the two poles of the cloud are distributed in horizontal layers with a vertical thickness L of neutral water-drops between them. The order of magnitude of

L is known from the lengths of lightning discharges to be 2 kilometres, and the critical value of the field, as we have seen, is 10,000 volts/cm. Thus the difference of potential between the poles is $10.000 \times 2 \times 10^5$ or 2×10^9 volts.

When a cloud-charge is conveyed to earth, or neutralised by another charge of opposite sign, by a lightning discharge, the energy dissipated is of the order of 2×10^{17} ergs or 5×10^9 calories; most of this is converted into heat along the path of the lightning flash. A cloud giving one flash every twenty seconds is dissipating electrical energy in the form of lightning at an average rate of 10^6 kilowatts. We have seen that other effects may account for an expenditure of energy at a rate considerably exceeding this.

§ 43. The Distribution and Origin of the Electric Charges

It is generally agreed that the seat of the generation of the electricity of a thundercloud lies within the cloud itself. Equal quantities of positive and negative electricity are produced and then separated by some agency into different regions, where they constitute the "poles" of the electrical machine. During the period in which the charges are built up the dissipation losses in these two regions may well be quite different and the quantities ultimately collected upon the two poles are not necessarily the same. The distribution of the charges has been the subject of a great deal of investigation, both practical and theoretical. At the present time two quite different views as to the distribution and the mechanism which gives rise to it, suggested by G. C. Simpson and by C. T. R. Wilson, are being examined.

According to Wilson the polarity of a thundercloud is what is called positive, the positive pole being higher up in the cloud than the negative one. The cloud is to be considered as made up of large negatively charged

drops and small positively charged droplets.[5] When a flash of lightning has just removed the free charges at the poles, the situation is momentarily that illustrated by Fig. 21 (*a*), in which the cloud as a whole is electrically neutral. Immediately afterwards, however, the faster rate of fall under gravity of the large negative drops leads to the development of free positive and negative charges at the top and bottom of the cloud, separated by a neutral region, as represented in Fig. 21 (*b*). Since

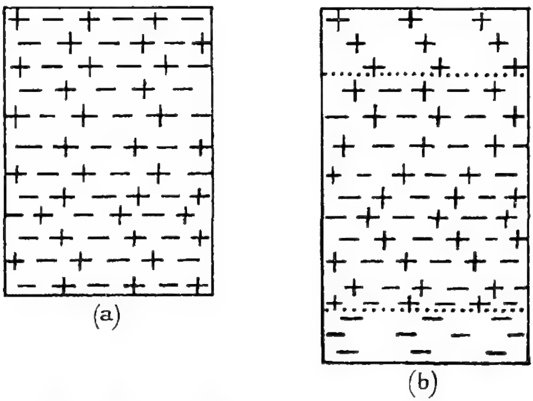

FIG. 21.—Distribution of thundercloud charges (C. T. R. Wilson).
(*a*) Just after a lightning flash.
(*b*) Just before a lightning flash.

the free poles attract each other, the rate of separation will become slower as the charges grow and will lead to a rate of fall of the larger drops which is just sufficient to maintain the charges at their existing values, in the face of the dissipating action of point discharge currents within and outside the cloud. Ultimately the field between the poles may become strong enough for a discharge to pass. If the charges are equal, this leaves the cloud in its initial state and the large drops in the neutral region once more separate under the action of

gravity. The relative rate of separation may be taken from Macky's results, already discussed, to be the terminal velocity of a drop 0·15 cm. in radius, which is about 6 metres per second. Relative to such a drop the much slower small droplets may be considered as stationary. The average rate at which the initial regeneration of the moment of the cloud takes place is 60/5 coulomb-kilometres per second, using the data given in § 42. From these figures Wilson [5] finds that the average total charge on the positive or negative carriers in the neutral part of the cloud is about 500 coulombs, occupying a volume of some 4 cubic kilometres. The charge per cubic centimetre of water on the larger negatively charged drops he finds to be less than 30 e.s.u., a result which we have seen (§ 36) to be in accord with direct measurements of the charge on rain from thunderstorms.

The mechanism suggested by Wilson as the cause of the opposite charges on large and small drops depends upon the presence in the cloud of numbers of slow ions, attached to Aitken nuclei and small water-drops. Since their mobility (§ 1) is about ·0003 cm./sec., the velocity with which these ions will move, even when the field attains its maximum value of 10,000 volts per metre, is only 3 cm./sec., which is less than that of a drop of water a few thousandths of a centimetre in radius falling under gravity. In the earliest stages of the development of the charged cloud, these may arise from natural sources, but later on they will be supplied in enormous numbers by brush-discharge from water-drops drawn out by the field into pointed forms.

Consider a large drop, say 1 mm. in radius, inside a cloud in which a strong downwardly directed field already exists. Owing to this field, the drop carries a negative polarisation charge upon its upper surface and a positive charge below (Fig. 22). As it falls downwards, it will interact with positive ions moving more slowly down and with negative ions moving up; encounters with other charged drops will be relatively rare. The

polarisation charge on the drop and the fact that its velocity, 590 cm./sec., is large compared with that of the slowly moving ions, will cause it to behave differently in its encounters with the two streams. At its under-surface a selective capture of negative ions will occur, for the negative ions which it meets will be attracted and the positive ions repelled by the positive polarisation charge there. A reverse process, a selective capture of positive ions, does not, however, take place at the upper surface, for here the drop is receding from the ions and the positive ions have been repelled away from the drop at the earlier encounter. Innumerable episodes of this kind will lead to the acquisition of a considerable resultant negative charge by the drop from the ion-streams, without seriously affecting the possibility of further captures. For example, a drop 1 mm. in radius falling in a field of 10,000 volts/cm. carries at each end a polarisation charge of ·25 e.s.u. and if it had collected a nett charge equivalent to the very reasonable value of 12 e.s.u. per cubic centimetre of water (§ 36), the negative ions captured would only increase the upper negative and decrease the lower positive charges by 10 per cent.

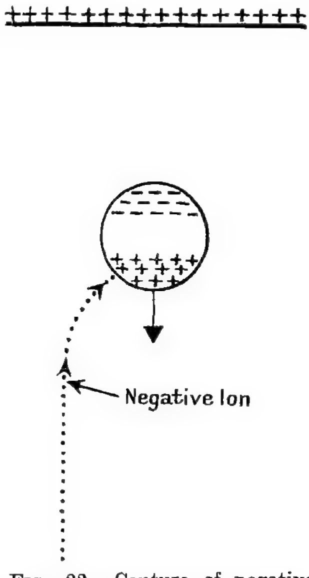

FIG. 22.—Capture of negative ions by large falling drops.

The tiny droplets in the cloud, which are falling with a velocity of the same order as that of the ion-streams,

would be unable to exercise the selective action described; they will therefore acquire a nett positive charge, the complement of the negative charge collected by the larger drops.

A complete account of this theory has not yet been given; it would involve consideration of the early stages in the building up of a charged cloud, the action of the normal potential gradient of fine weather in determining the direction of the field to be generated, and the initial building up with the aid of natural ions.

A very different view of the origin and distribution of thundercloud charges has been developed by Simpson,[6] who considers the generation of the electricity to be a consequence of the disruption of rain-drops. Lenard showed in 1892 that when pure water splashes against a solid obstacle the water becomes positively charged, while a complementary negative charge is given to the air. Simpson in 1908 extended these experiments to the breaking of drops in an ascending current of air and obtained the same result, and quantitative values for the charge developed in the process. In extending this observation to conditions within a thundercloud, account has to be taken of the fact that a water-drop of radius greater than ·25 cm. becomes flattened out and unstable when it falls through the air, with the result that it breaks up into a number of smaller drops. Since the terminal velocity of fall for a drop ·25 cm. in radius is 8 metres per second, it follows that no drops of water can ever fall downwards through an ascending current of air whose vertical velocity component exceeds 8 metres/sec. All drops of water under such circumstances will be carried upwards.

The meteorological conditions inside a thunderstorm of the heat-type are represented by Simpson as in Fig. 23; the continuous lines represent stream-lines of the air, and their distance apart is made inversely proportional to the wind-velocity at each point. The air enters the cloud from the right and passes upwards into the front of the cloud. The vertical component

of the velocity of the wind is taken to reach a value exceeding 8 metres per second within the oval unshaded region. No water, as we have seen, can pass vertically downwards through this region. The paths of the larger drops, as they fall through the cloud, are represented by the dotted lines. On the right-hand half of the figure these drops are deflected to the left by the air-stream. Above the oval region, therefore, there will be a considerable accumulation of water, and it is here that the larger drops will form and break up on account of instability, giving rise to electric charge. The largest drops, ·25 cm. in radius,* will be falling fast

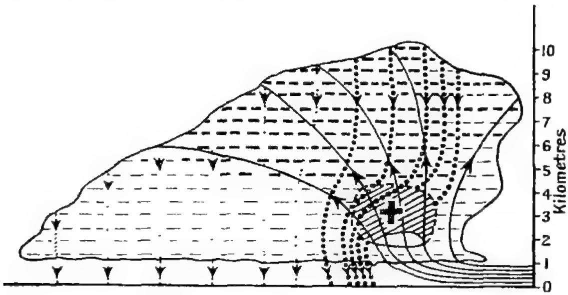

Fig. 23.—Thundercloud electrification (Simpson).

enough to penetrate to just above the forbidden oval, where they will be broken into pieces by the upward-moving air. The fragments, now positively charged, re-form drops and fall back again, and the negative charge given to the rising air will be absorbed by small droplets and carried to the main body of the cloud. The region in which the process of generation of electricity occurs is indicated by the heavily shaded area above the oval.

Simpson shows that a reasonable estimate for the average mass of water in the shaded region is 3×10^{11} grams, the dimensions of the region being of the order

* The theory in this form was put forward before Macky's results (§ 42) were published.

of a kilometre. If at any one time half this water is in the form of drops large enough to be broken up, the total number of drops available for breaking is $2{\cdot}2 \times 10^{12}$. Since the laboratory experiments showed that a drop of radius ·25 cm, when broken, gives rise to $5{\cdot}10^{-3}$ e.s.u. of charge, the quantity developed if every available drop broke once would be 3·6 coulombs. The drops would have to break six times to produce the 20 coulombs on the poles of the cloud.

To decide between these two views on the charge distribution, various experimental tests have been applied to thunderclouds, the description of which is beyond the scope of this book. For the discussions on this question, the reader may be referred to the references at the end of the chapter [2, 3, 6, 7, 8, 9]. It is not impossible that both the Simpson and the Wilson mechanism may be at work in a thundercloud at one and the same time.

§ 44. Lightning

Studies of the lightning discharge are of considerable interest in relation to the problems just discussed; they also furnish valuable material for the study of the spark discharge in air at ordinary pressures, over the mechanism of which general agreement has not yet been reached.

When photographs of the lightning channel taken on a moving camera are examined, they show that many apparently continuous discharges are made up of a number of separate strokes, separated by short periods in which faintly luminous channels sometimes continue. These discontinuities are often apparent to the eye as flicker of the flash of lightning. Though the whole flash often occupies more than a second, the component strokes are generally over in a few thousandths of a second. After each such stroke, the ionisation in places along its track persists sufficiently long to provide the succeeding stroke with a ready-made conducting path. Fig. 24 (after Walter) [10] illustrates how the apparent

branching of a flash observed by the eye (24 (*a*)) is often the result of the superposition of a number of separate strokes, whose tracks coincide in part. Fig. 24 (*b*) is from a photograph of the same flash as (*a*), taken on a rotating camera.

These discontinuous strokes must arise from discontinuities in the rate at which the thundercloud can supply electric charge to the end of the channel. Since the cloud itself, though charged, is only a conductor by virtue of the ions produced by point-discharge from drawn-out water-drops, the conditions within the cloud, during the passage of a discharge, must be very complex.[11]

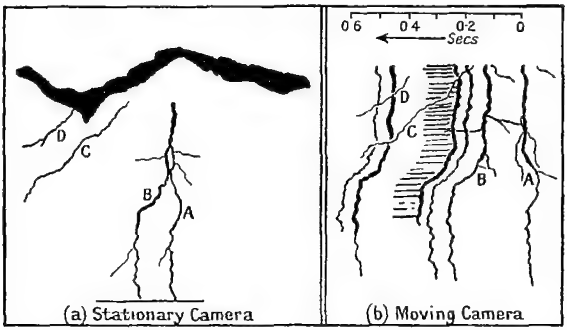

FIG. 24.—Component lightning strokes (Walter).

The field-change apparatus of Wilson, in the form described, does not usually yield information as to the quantities of electricity conveyed by these component strokes, the recording arrangement being too slow to record them separately. The results given in § 41 refer to the total integrated field-change and so to the total charge transferred by the flash.

§ 45. OSCILLATIONS IN THE LIGHTNING STROKE

Much attention has been paid to the question as to whether a lightning discharge is oscillatory or aperiodic,

and several writers have examined the applicability of the well-known theoretical condition for aperiodicity, $R^2 > 4L/C$, where R is the total resistance of the circuit, C the capacity and L the inductance. The thunder-cloud itself is considered by some as a condenser plate whose capacity to earth can be estimated from its dimensions, by others as a non-conductor. Simpson,[11] who takes the latter view, has pointed out that the conducting channel of the flash, with its ramifications and tributaries within the cloud, is a conducting system possessing distributed capacity and self-inductance. He has shown that a straight lightning channel 2 km. high and 5 cm. in diameter will be capable of oscillation if its total resistance is less than 2000 ohms, or 1 ohm/metre. A potential difference of 10^9 volts applied to a channel of this resistance would give rise to a current of 500,000 amperes; there is evidence that momentary currents as large as this do actually flow.

On this view the oscillations may be expected to take the form of ripples superimposed upon the main current, without reversing its direction of flow. Such ripples have been observed by several workers (Appleton, Watt and Herd [7]; Norinder [14]; Mathias [20]) who have used the cathode-ray oscillograph to record the field-changes due to distant lightning flashes. The main effect is a unidirectional pulse which rises to a maximum and then falls to zero. On this are superimposed ripples which have a frequency corresponding to a wave-length in air of about 30 km. If the lightning channel be regarded as a gigantic wireless aerial 3 km. high, connected to earth at its base, its natural wave-length would be 12 km. Since the actual channel has extensive branches within the cloud, the agreement with the observed wave-length as regards order of magnitude can be considered satisfactory.

§ 46. The Electrical Effects of Direct Strokes

We have seen (§ 42) that the potential of a thunder-cloud pole is about 10^9 volts. An elevated and insulated

conducting system when struck by lightning may be expected to attain a considerable fraction of this potential. Experiments of this nature have been made by Brasch and Lange,[12] who used a wire stretched across an Alpine valley and insulated at each end. They have reported potentials of about 10,000,000 volts, giving sparks to earth 18 metres long. Pittman and Torok,[13] using a cathode-ray oscillograph connected to an overhead power line, have recorded a potential of 5,000,000 volts caused by a travelling wave from a lightning stroke which hit the line four miles away.

The time occupied by a lightning stroke is an important quantity in connection with the current in the discharge. Measurements by Norinder,[14] Mathias[20] and Watson-Watt[7] agree in giving values of the order of ·001 second for the duration of single strokes. Since we know from the work of Wilson that the average quantity of electricity carried by the stroke is about 20 coulombs, the average value of the current must be of the order of 20,000 amps. One would, however, expect that for very short periods of time the discharge current is much greater than this average. Lightning strokes to the supporting towers of power transmission lines have been found to carry currents as high as 250,000 amps.

The question of the diameter of the discharge channel has been the subject of laboratory experiments by Petersen,[15] who has found that powerful spark discharges carrying currents of the order of 60,000 amps. have a constant current-density of 10 amps. per square millimetre. If this result may be extrapolated to currents of 200,000 amps. it would yield 16 cm. for the diameter of the channel of a powerful lightning flash. The fulgurites formed by the fusion of sand struck by lightning have a diameter of some 5 cm., and Petersen reports that in his heavy current discharges the diameter of the channel at the electrodes contracted to one-third of its value elsewhere.

A simple and valuable instrument for the study of direct lightning strokes has been developed in America

under the name of the Klydonograph. This is an adaptation of the well-known Lichtenberg figure experiment and operates as a recording voltmeter for high potential differences. The positive electrode (Fig. 25) takes the form of a rod A, touching the emulsion of a photographic plate or film. Below this is a thin layer of varnished paper insulation resting upon a flat metal plate connected to earth. When A is raised to a high positive potential a discharge passes along the photographic emulsion, which on development shows as a circular patch or figure. It has been found that the relation between the diameter of this figure and the applied voltage is a linear one up to potentials of 25,000 volts. Above this the instrument must be used with a potential divider, tapped off, for example, from a string of insulators connected between the elevated conductor and the earth. To measure negative potentials the connections must be reversed, the rod A earthed and the plate connected to the high potential side. For lightning recording, two oppositely connected klydonographs have been joined to the transmission lines under examination.

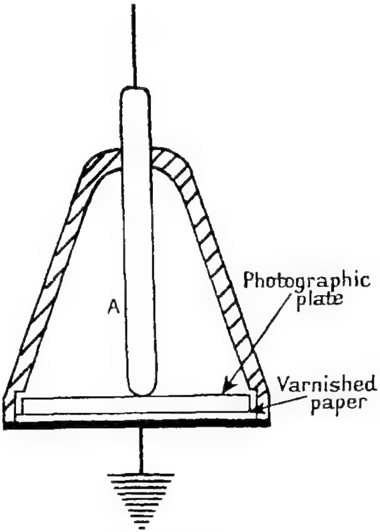

Fig. 25. — Klydonograph arrangement for recording potentials in lightning strokes.

Lewis and Foust [16] have recently published the results

of some years' investigations with 1500 klydonographs connected to the supporting towers of transmission lines. One hundred direct strokes were recorded and all these conveyed negative charge from cloud to ground. This result is of interest in connection with the problems of the maintenance of the earth's charge (§ 38), and of the distribution of electricity in the thundercloud.

§ 47. The Mechanism of the Discharge

Two discussions of the actual mechanism of the lightning discharge have been given. Dorsey[17] considers that the principal factor is the movement of a stream of fast electrons from the negative pole, the ions necessary for the subsequent passage of the main discharge arising from the Townsend collision process. According to Simpson,[18] however, the discharge proceeds away from the positive pole; the slower-moving positive ions, on this view, constitute the tip or spear-head of the discharge as it develops, and the negative ions produced at the tip flow into the positive pole along the channel.

Light may be thrown on this important question, which is directly connected with the vexed problem of the mechanism of the electric spark between metal electrodes, by photographs taken with a camera moving sufficiently rapidly to show the manner in which the first stroke in a lightning flash proceeds. C. V. Boys[19] has developed a camera of this type and has published one such photograph. It shows a flash beginning at a point some 500 metres above the earth and developing its track simultaneously above and below.

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